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Lane, R. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 35c.)

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JUNE 1956

THE GRANITES OF TRANCOSO (PORTUGAL): A STUDY IN MICROCLINIZATION

L. J. G. SCHERMERHORN

ABSTRACT. At Trancoso, Portugal a massif of medium-fine granite is embedded in a younger coarsely porphyritic granite. The microcline megacrysts of the porphyritic granite exhibit zoning, orientation of inclusions, and hourglass structure. The older granite consists of a granodiorite which in many places has been extensively microclinized to granite under the influence of the porphyritic granite. Textural and quantitative evidence is given. All plagioclase in both granites and in xenoliths is albite-oligoclase. This seems to be due to secondary decalcification of originally more calcic plagioclase. The plagioclase shows albitic outer zones where in contact with microcline. Both granites carry abundant muscovite and occasional sillimanite as metasomatic minerals.

INTRODUCTION

Coarsely porphyritic granite has by far the greatest distribution among the granites of northern Portugal. Fine- to medium-grained granites of various types make up the remainder.

The area around the ancient walled town of Trancoso permits study of the relations between an older granodiorite and the encasing younger, porphyritic granite. The composition of the older rock, the Trancoso granite, ranges from granodioritic to granitic, under the influence of microclinization by the porphyritic granite. A special feature of the two granites contact is a profusion of pegmatites.

A much-weathered quartz schist crops out about 3 km southeast of Trancoso, most probably belonging to the pre-Ordovician Complex, a monotonous formation of generally metamorphic pelites and psammites unconformably underlying the Ordovician of northern Portugal. The schistosity of these rocks strikes 60° east of north, with approximately vertical dip. The porphyritic granite to the north occasionally carries angular hornfels inclusions derived from the schist.

Mafic dikes, varying in width from less than 10 cm to 5 meters are rather common. They consist of a very fine grayish-green rock with occasional feldspar phenocrysts. The dikes are perpendicular or nearly so and strike predominantly NNE-SSW or about NW-SE.

Important NNE-SSW faults cross the area.

The accompanying sketch map (fig. 1) roughly indicates the distribution of the various rock types.

Optic axial angles given were obtained by direct measurement on the universal stage; anorthite contents of plagioclase were determined by means of Fedorov migration curves.

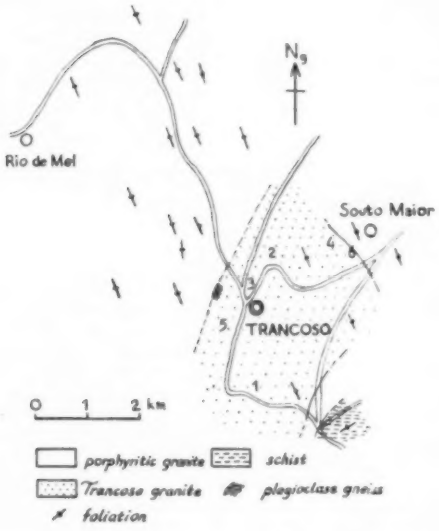

Fig. 1. Geological sketch map of the Trancoso area. The numbers indicate the localities of the specimens of which modal analyses are given in table 1.

THE PORPHYRITIC GRANITE

The porphyritic granite is to serve as reference unit, both petrographically and geologically. Its influence on the older granodiorite has been profound. The emplacement of the porphyritic granite took place after the second Hercynian phase in northern Portugal which sharply compressed the open folds of the first phase in which the upper Carboniferous was deposited. This granite may be observed crosscutting and metamorphosing Stephanian psammites northeast of Sátão, 33 km due west of Trancoso. Between the metasediments of Sátão and the Trancoso granite all is porphyritic granite. The block-faulted nature of this area is brought out by alternating stretches of deeply weathered plains and rocky uplands, divided by northeasterly- to north-northeasterly-striking faults. The porphyritic granite typically weathers into enormous rounded boulders or woolsacks (fig. 2), from which the more resistent microcline megacrysts protrude. The rock consists of abundant potash feldspar megacrysts (tabular on 010 and displaying the forms {010}, {001}, {110} and {101}) in a coarse matrix of feldspar, quartz, biotite, and white mica. The megacrysts may reach 4 cm in length by 1 cm across. Biotite and silvery-white mica frequently form six-sided flakes up to 4 mm in diameter. The quartz is in round and irregular grains which attain 10 mm in size. The granite may carry more muscovite than biotite. Although the texture of the rock is uniform throughout the area shown in the map, the dimensions of the constituents may vary somewhat. The coarsely porphyritic granite west of Trancoso grades into a less coarse variety towards Rio de Mel in the northwest. The passage is continuous and very gradual. In the latter variety the sizes of the constituents are somewhat reduced, on a scale of about 3:4 (compare fig. 3).

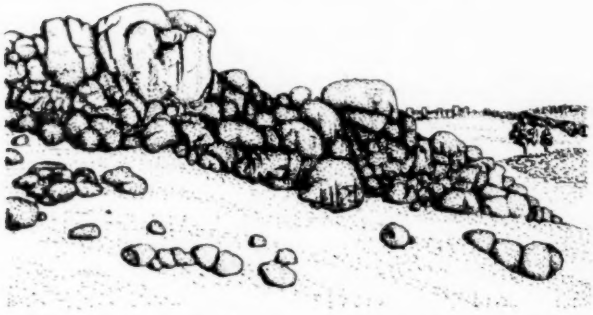

Fig. 2. Woolsacks of porphyritic granite, northwest of Trancoso. In the background the walls of Trancoso.

The porphyritic granite generally shows a planar parallelism, or *foliation*, of the megacrysts. The inclination of the foliation is always about vertical (± 10°); the strike varies between N 140 and N 180 (azimuthal reading), averaging N 160. In places a faint *lineation* is shown, a linear parallelism of the longest dimensions (c-axes) of the megacrysts. It is about vertical (± 20°). Foliation and lineation are considered due to regional stress after emplacement, since the megacrysts developed in a more or less stationary environment after intrusion was completed. This is more fully discussed in Schermerhorn (1956), in which the petrogenesis of a porphyritic granite at Cete, a hundred kilometers to the northwest, is described.

The primary accessories consist predominantly of *zircon*, *apatite*, and *ilmenite* (frequently replaced by leucoxene). They are idiomorphic and occur chiefly in biotite (preferential distribution). Some apatite crystals may reach diameters of over 0.5 mm.

There is a slight amount of what seems to be *early muscovite*, in small idiomorphic flakes included in biotite.

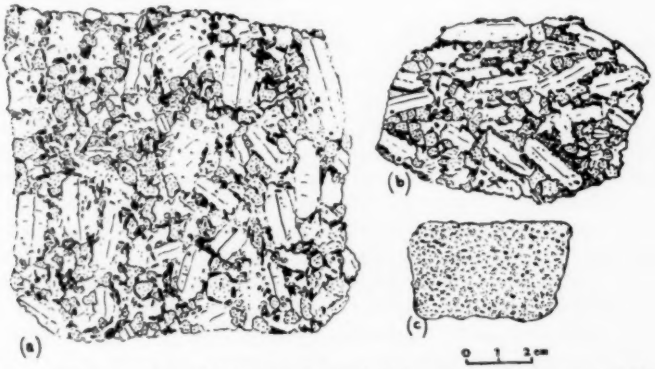

Fig. 3. Plane surfaces of (a) coarsely porphyritic granite, 1½ km west of Trancoso; (b) less coarse porphyritic granite, halfway between Trancoso and Rio de Mel; (c) granodiorite, 200 meters north of Trancoso. White: feldspar; black: mica; stippled: quartz.

Biotite is pleochroic from deep red brown to straw yellow and carries abundant pleochroic haloes around zircon, sphene, or granules too small to identify, and occasionally around apatite. Biotite was the earliest of the main constituents to separate; it may be enclosed and/or corroded by plagioclase, microcline, quartz, and late muscovite. Preferential distribution of the primary accessories in the biotite points to its magmatic growth. The theory of this is given in Schermerhorn (1956); briefly, it is argued that since the primary accessories largely separated prior to biotite, their concentration in the mica was effected during the latter's crystallization period, in a fluid environment. It is thought that this concentration was made possible by internal motion of the intruding magma so that most of the earlier accessory minerals came successively into contact with growing biotite flakes, adhered to their surfaces (surface tension), and were incorporated. The latest biotite to form is represented by minute rounded flakes in quartz III (main quartz), which do not carry accessories. The biotite may be altered into a penninite-like chlorite, with anomalous blue interference colors. An intermediate stage is frequently represented by a green mica with somewhat lower birefringence than biotite. Both the green mica and the chlorite may show development of sagenite. Decomposition of biotite is often accompanied by segregation of sphene granules and aggregates (along the cleavage of the mica) with pleochroic haloes.

Quartz displays the following six different habits: quartz I—bipyramids; quartz II—inclusions in microcline megacrysts with concave or partly straight boundaries; quartz III—main quartz, forming the large grains in the matrix; quartz IV—drop-quartz; quartz V—small irregular bodies; quartz VI—garlands. Quartz I is a separate generation; quartz II, III and IV are considered a single generation, starting separation in the magmatic stage and continuing into the metasomatic stage; quartz V and VI seem to be still later (compare Schermerhorn, 1956).

Quartz I forms small bipyramids of original high quartz in feldspar. The other quartz varieties are posterior to feldspar.

Plagioclase builds hypidiomorphic, generally albite-twinned crystals, many of which are sericitized and kaolinized. A little fluorite is present among the alteration products. Plagioclase is corroded by microcline (fig. 4), quartz, and white mica. The large crystals in the matrix (up to 8 mm in length) consist of *albite-oligoclase* (An_{9-15}, averaging An_{12}). The plagioclase inclusions in microcline megacrysts (up to 2.5 mm in length) are slightly zoned, showing albitic rims (An_{0-5}) grading into somewhat more calcic cores (up to albite-oligoclase (An_{12}) in grains longer than about 1 mm). The cores are sericitized, the albitic rims are clear. Where the albite-oligoclase of the groundmass is in contact with microcline, it also shows a gradual decalcification of the outer zones, up to albite. These albitic rims follow corrosion outlines and are absent against adjacent minerals other than microcline. Myrmekite is occasionally developed in plagioclase corroded by microcline.

The potash feldspar is *microcline microperthite*, building roughly euhedral megacrysts and irregular grains in the matrix. $2V\alpha$ varies from 78° to 86°. Extinction angles are 15°-16° on 001 and 6°-7° on 010. Cross-hatch twinning is often present. It may be coarse and patchily distributed in un-

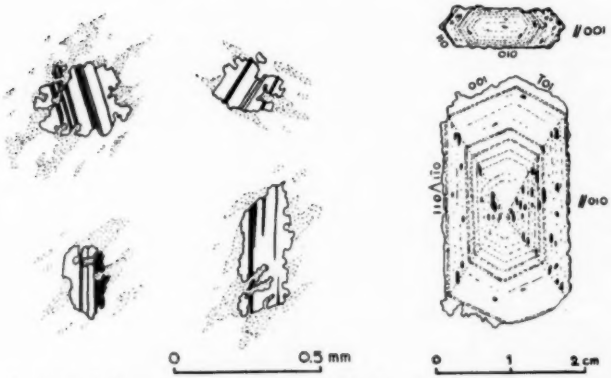

Fig. 4. Corroded unoriented plagioclase inclusions in microcline megacrysts. Perthitic plagioclase is stippled.

Fig. 5. Microcline megacryst in porphyritic granite. Above, section parallel to 001; below, section parallel to 010. Note euhedral zoning (narrow white opaque bands alternating with wider translucent zones), irregular border and hourglass structure of biotite inclusions (plagioclase and quartz inclusions not shown). The flakes of biotite are mostly parallel to microcline {110}; some are oriented on {001} or {10$\bar{1}$} in the upper and lower parts of the hourglass pattern.

twinned undulant microcline. Perthite is mainly of the vein and patch type; film perthite may also be present. The perthitic plagioclase may reach An_{10} in the patches but is generally more sodic in composition.

The *orientation of inclusions* of older minerals in microcline is definitely a magmatic texture. Orientation is in three types: on structural directions in the host, in zones, and in hourglass pattern. The three types are frequently combined. The oriented inclusions, up to a few millimeters in size, consist of the tabular minerals, biotite, and plagioclase. These are older than the microcline as proved by corrosion relations (compare figs. 4, 6). Any distinct orientation of these inclusions with respect to the structure of their host is therefore due to influences exerted by the host during its crystallization, and must of necessity have occurred in a fluid environment permitting rotation of existing crystals. The biotite and plagioclase inclusions are generally oriented on the {010}, {001}, {110}, and {10$\bar{1}$} planes of the microcline megacrysts in which they occur. {110} seems to exert the strongest orientational control in the porphyritic granite in question, since most of the inclusions exhibit 110-orientation (parallelism of biotite 001 and plagioclase 010 to microcline {110}). This is a dimensional orientation; the plagioclase inclusions may in addition show a structural orientation. In this case they exhibit parallelism not only of the 010-planes to a crystallographic direction in the microcline but also of some structural direction, generally the c-axis, to a direction in the host. Oriented plagioclases are in general less corroded than unoriented ones. Figure 6 shows part of a megacryst, cut parallel to 010, with oriented inclusions. One set of plagioclase inclusions in this megacryst is oriented on microcline 001; these show sharp 010-twinning lamellae. The

other set is oriented on microcline 110, exhibiting undefined boundaries between the lamellae, cut at a low angle. The biotite flakes are also oriented. Fourteen measurements on plagioclases of the first set yielded $010_{plag} \wedge 001_{micr}$ angles varying from 1 to 9°; 19 measurements on the second set indicated $010_{plag} \wedge 1\bar{1}0_{micr}$ angles between 0 and 10°.

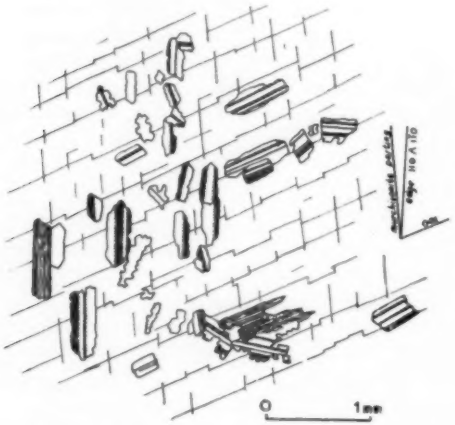

Fig. 6. Oriented albite inclusions in a microcline megacryst cut parallel to 010. There are two directions of orientation: plagioclase 010 parallel to microcline 001, or 110. Note the two oriented biotite flakes, one with apatite inclusions.

The oriented plagioclases and biotites frequently display a distinct zonal arrangement. A related directional intergrowth is formed by an *hourglass structure*, as far as I am aware hitherto unrecorded from feldspar. It combines the orientational arrangement of inclusions described in the preceding paragraph with a distribution in an hourglass pattern. This is illustrated in figure 5, where the orientation of the biotite and plagioclase inclusions is principally on {110} in the microcline megacryst. The section parallel to 010 passes approximately through the center of the megacryst. In the figure only the biotite flakes have been drawn, for the sake of clearness; most of them lie parallel to the nearest {110} plane (compare the 001-section, in fig. 5). The hourglass structure is evidenced by concentration of the biotites in two triangular areas between the megacryst center and respectively the 110-1$\bar{1}$0 and 1$\bar{1}$0-1$\bar{1}$0 edges. It is also shown by the microcline being less translucent in these areas. The 001-section passes at some distance from the megacryst center so that it does not exhibit the pattern to most advantage. The spaces containing the biotite are shaped as two pyramids along the a-axis, with their apices facing in the center (evidently the crystallization nucleus) of the microcline host. The structure, which I have often observed in northern Portuguese granites, is rather well developed in the porphyritic granite around Trancoso.

The orientation of the inclusions is inferred to have been effected by crowding aside of tabular older minerals by the successive growth surfaces of the microcline megacrysts. The hourglass structure moreover depends obviously on variations in the crystallization force (or rather ability to crowd aside older crystals) of the potash feldspar according to the different directions in its lattice. In 010 it seems to have been greater parallel to c than normal to

c, so that pre-existing crystals were pushed aside, to the 110 planes of the growing megacrysts, clearing the area of the c-axis (along which the megacrysts are invariably longest). Incidentally, these orientational textures bear out the primary nature of the zoning displayed by the same megacrysts. The *zonal structure* of the megacrysts is best observed macroscopically (fig. 5). Under the microscope it is vaguely evidenced by narrow bands somewhat richer in perthitic plagioclase and more kaolinized than the wider zones with which they alternate. It may be regarded as a kind of oscillatory zoning brought out by alternate shells of different behavior, possibly on account of somewhat varying sodium content. This zoning is a primary structure, its idiomorphism indicating development by zonal accretion in a magmatic environment.

The features described indicate growth in a fluid environment.

The potash feldspar megacrysts are thus shown to have largely crystallized from a magmatic fluid. They ended crystallization by a final metasomatic stage, as shown by their irregular corrosive borders.

An exhaustive study of potash feldspar features indicating magmatic growth has recently been published by Frasl (1954), concerning the potash feldspar megacrysts of some porphyritic granites and gneisses in Austria. According to Frasl, both orientation of inclusions and zoning (he does not mention hourglass structure although his figure 3 seems to exhibit it) may be taken to prove growth from magmatic fluids at high temperatures. He also discusses the evidence presented by habit, twinning, alteration, red coloration by hematite, optical orientation and type of perthitic structure; their value as yet seems relative. In my opinion it seems necessary to add that to the possibility of rotation of inclusions in a fluid phase it is irrelevant whether this fluid is truly magmatic or has formed, by anatexis, in and from solid rock. It will be shown below that orientation of biotite and plagioclase may occur in the new microcline of the Trancoso granite which developed in originally solid rock.

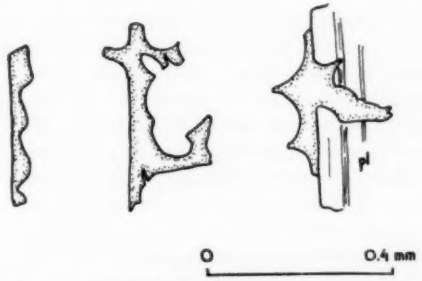

Fig. 7. Shapes of quartz II inclusions in microcline megacrysts.

Quartz II forms occasional inclusions in microcline megacrysts, with concave or partly straight outlines (fig. 7). Its straight boundaries are molded on plagioclase or biotite inclusions in the megacrysts or follow structural planes in the microcline (ancient growth surfaces). Quartz II is therefore considered to have crystallized, together with that part of the surrounding

microcline towards which it is concave, from occasional pockets of residual magma enclosed by the growing megacrysts.

Quartz III is abundant, in large grains corroding biotite, plagioclase, and microcline. Its formation started in the magmatic stage to continue into the metasomatic stage. Magmatic crystallization is suggested by textures such as may be discerned in figures 3a and b, in which flakes of biotite are tangentially arranged around large rounded grains of quartz. Since biotite is older than quartz III, as borne out by corrosion relations, this texture may be inferred to have originated by a crowding aside of the mica flakes by growing quartz, until they abutted against surrounding minerals. Slightly bent plates of biotite, appressed between quartz III and feldspar, are occasionally observed. Thus the quartz crystallized interstitially in a rather loose mesh of biotite, plagioclase, and microcline. It continued metasomatically, as indicated by its replacive attitude toward the older minerals. It is concluded that after quartz III had separated, the magma had entirely solidified, except for a certain amount of rest-fluids present in, or introduced into, the intergranular spaces. Minerals posterior to quartz III and transecting grain boundaries are therefore metasomatic. The quartz commonly exhibits undulatory extinction, often with some granulation. This is thought to be due to the post-granitic block-faulting of the area.

Quartz IV, or drop-quartz, is sparingly present as small rounded bodies metasomatically formed in feldspar crystals near the margins. Quartz IV grains tend to occur in groups, frequently in optical continuity, mutually and with quartz III crystals nearby.

Quartz V is occasionally found in plagioclase, as small irregular bodies which frequently show stepped or frayed outlines oriented on the host structure.

Quartz VI is sparse. It forms narrow veinlets or garlands in microcline, following the cleavage directions of the host.

Sillimanite is uncommon. It has been encountered in a few slices, at the contact with Trancoso granite. The sillimanite is associated with late muscovite in which it occurs as bundles of needles. It generally shows fibrolite habit, but may build crystals up to 0.15 mm in width. Fibrolite trains in muscovite may in places continue into adjacent quartz III crystals. They are therefore posterior to quartz III and metasomatic. Sillimanite may attack biotite, starting at the borders of the mica and replacing it by a mat of felted fibrolite in which biotite is reduced to vague still pleochroic remnants before disappearing altogether. Figure 8 shows fibrolite developed at the expense of biotite, surrounded by late muscovite.

Muscovite is a late metasomatic mineral, apart from a slight amount of early white mica included in biotite. It forms irregular crystals, in places with a tendency to idioblastic outlines, but generally with ragged borders. It replaces biotite, plagioclase, microcline and, to less extent, quartz. The mineral has $-2V$ about 37° and a birefringence of 0.033-0.035.

Attention may be drawn to some differences between this granite and the Cete granite which probably belongs to the same batholith (Schermerhorn, 1956). Apart from the occasional presence of sillimanite the most important

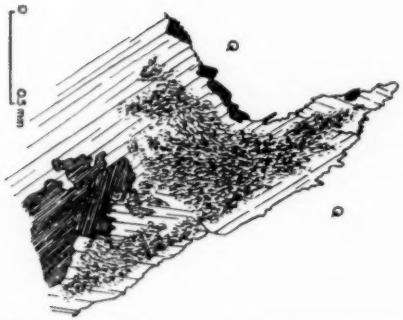

Fig. 8. A felted mass of fibrolite resulting from near-complete replacement of biotite, surrounded by muscovite. Porphyritic granite.

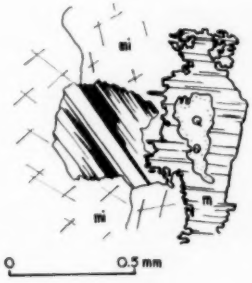

Fig. 9. Plagioclase corroded by microcline and muscovite. Note stepped corrosion outlines and tapering twin lamellae. Trancoso granodiorite.

difference is the uniform albite-oligoclase composition of the plagioclase. It will be argued below that this feature is secondary. Minor differences lie in the relative amounts of the late-stage minerals. In the granite in question there is more secondary muscovite and less late quartz than in the granite at Cete (there seems to be more quartz III). These differences originated in the metasomatic stage. The magmatic crystallization of both granites is considered to have been very similar, if not identical, slight divergences developing afterwards.

THE TRANCOSO GRANITE

The massif of Trancoso, a few kilometers across, consists of a medium-fine granite to granodiorite (fig. 3c). The difference in weathering between the huge woolsacks of the porphyritic granite and the small angular blocks of the granodiorite is conspicuous.

The rock is non-porphyritic except for occasional disseminated microcline megacrysts near the contacts with porphyritic granite or with pegmatites. It consists of white feldspar and quartz, with scattered small plates of biotite and very irregular flakes of white mica. Subparallel arrangement of biotite may occasionally lend a faintly gneissose appearance to the granodioritic varieties; it is probably a primary directional texture, destroyed during the subsequent microclinization of the rock. The muscovite does not exhibit any orientation. Where the rock shows extensive development of new microcline, in tabular grains with euhedral tendencies, these may display a foliation (planar parallelism) of the same north-northwesterly strike and vertical dip as the foliation in the porphyritic granite. Also, the amount of this microcline and of disseminated megacrysts increases conspicuously toward the porphyritic granite or around pegmatites. Thus a microclinization conditioned by the porphyritic granite and under the same stress is already evidenced in the field.

The primary accessories, *zircon* and *apatite*, are present in smaller amount than in the porphyritic granite. There is little or no ore (leucoxene,

presumably pseudomorphic after ilmenite). Zircon builds minute prisms, preferentially distributed in biotite, or in muscovite replacing biotite. Apatite (up to 0.8 mm long) occurs mainly in biotite but may also be met with in other constituents.

Biotite is pleochroic from dark reddish-brown to straw yellow and carries abundant pleochroic haloes. It is corroded by all subsequent minerals. Chloritization, with sagenite and/or sphene release, is of the same type as in the porphyritic granite.

Plagioclase is in grains up to 2.3 mm in length, invariably showing multiple twinning (apparently Albite Law twinning), occasionally combined with Carlsbad twinning. Rarely a few pericline lamellae occur. The plagioclase generally carries fine alteration products consisting of sericite and kaolinite. The plagioclase is not zoned; its composition is very constantly that of albite-oligoclase (An_{9-14}, averaging An_{11}). In the granodioritic varieties a very narrow border zone (max. 0.05 mm) is generally decalcified to albite (An_{2-5}) at contacts with potash feldspar. This thin sodic zone closely follows the plagioclase outlines as corroded by microcline, abruptly ceasing where the plagioclase no longer adjoins microcline. These decalcified rims have therefore formed by a secondary process connected with corrosion by potash feldspar. In the Trancoso granodiorite it is characterized by a very abrupt change of extinction when passing from the albite-oligoclase into its albitic border zone. In the more granitic varieties the sodic rims widen concomitantly with increasing microclinization, and pass into gradual borders such as occur in the porphyritic granite. This also illustrates the presumably genetic connection between this secondary decalcification to albite and the development of microcline.

Plagioclase is replaced by later quartz (figs. 10, 11), microcline (figs. 9, 10, 12, 13), and muscovite (fig. 9). Frayed and stepped corrosion outlines, following plagioclase structure, are not uncommon (fig. 9). It is occasionally

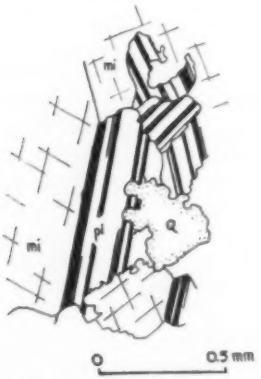

Fig. 10. Microcline corroding and molding plagioclase. Note lobate quartz. Trancoso granodiorite.

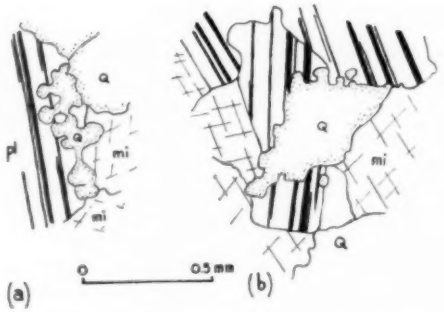

Fig. 11. Lobate quartz, corroding plagioclase and microcline. Note interstitial attitude of quartz in (*a*). Trancoso granodiorite.

observed that the polysynthetic 010-lamellae of plagioclase widen towards corrosion boundaries. This may indicate that the twinning of the plagioclase, at least in these cases, took place after corrosion. Myrmekite is occasionally developed in plagioclase corroded by microcline. Cataclastic effects are noted in places (fig. 12). Grains of plagioclase may be broken and the fragments slightly displaced. The twinning lamellae may vary in the fragments which points to twinning as a secondary process.

TABLE 1
Modes of Trancoso Granite (vol %)

	1	2	3	4	5	6
Microcline-perthite	12.5	24.5	22.9	29.3	27.9	29.9
Plagioclase	44.0	34.2	29.8	25.7	21.1	20.0
Quartz	29.0	31.3	27.9	31.6	33.4	33.0
Muscovite	9.5	5.2	11.8	5.6	10.0	9.9
Biotite, accessories	5.0	4.8	6.6	7.8	7.6	5.1
Sillimanite	—	—	1.0	tr.	—	2.1

Location of specimens given in figure 1.

Microcline occurs in varying amounts. The rock grades from granodiorite through adamellite to granite, according to Johannsen's classification (1932). This is caused by the development of microcline, principally at the expense of plagioclase, as shown in the modes in the accompanying table. The modal analyses were made by means of a Chayes point counter (Chayes, 1949). The modes are meant to demonstrate the variability of the plagioclase: microcline ratio in the massif as a whole. The microclinization is considered a secondary process brought about by contact action of the younger porphyritic granite, as stated below; it is most extensively developed near the contacts with the younger granite or with its pegmatites. Occasional disseminated microcline megacrysts occur near these contacts; here they may reach a density of one per square decimeter outcropping Trancoso granite.

In the non-porphyritic microcline two transitional varieties may be distinguished by means of perthitization and optic axial angle. They will be called "old" and "new" microcline. The former occurs in the more granodioritic types, giving way to the latter in the granitic types. The porphyritic microcline belongs to the latter variety.

The old microcline forms anhedral grains up to 2 mm in size in granodiorite. It may mold itself on euhedral outlines of biotite and plagioclase (without replacement) or may corrode them (fig. 10). It is itself attacked by quartz and muscovite. A shadowy undulatory extinction is common. The optic axial angle, $2V\alpha$, in old microcline is about 65° (mean of 22 determinations, from 60° to 69°). Film perthite is abundant and always present. It is formed by extremely thin films of albitic plagioclase (up to 0.0008 mm in width and reaching 0.05 mm in diameter) oriented on the murchisonite direction in the potash feldspar host. The film perthite does not grade into

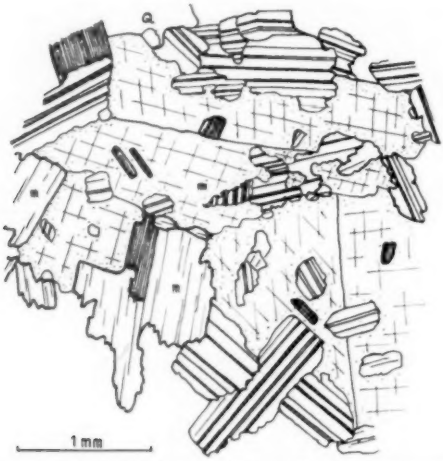

Fig. 12. Fabric of microclinized granodiorite, showing two crystals of new microcline (mi) corroding plagioclase, with flakes of muscovite (m). Note cataclastic plagioclase above. Trancoso granite.

vein or bleb perthite. Where albitic blebs or veins appear, the film perthite around them disappears.

The new microcline has an optic axial angle of about 75° (mean of 17 determinations, from 70 to 80°). It is slightly perthitic, exhibiting small irregular albitic blebs and veins.

Cross-hatch twinning is but faintly developed in both varieties of microcline, as a wavy indistinct grid, and is frequently absent, especially in the old microcline. The microcline is generally but slightly kaolinized, much less so than primary or perthitic plagioclase.

The development of the new microcline may be observed in hand specimens, since it tends to form roughly euhedral grains which may occasionally reach 8 mm in length. Carlsbad twinning is almost invariably present. The new microcline replaces biotite and plagioclase and is corroded by quartz and muscovite. All transitions occur, from microcline interstitial to biotite and plagioclase grains, or partly surrounding and corroding plagioclase, to large more or less euhedral plates of microcline. Biotite and plagioclase inclusions in new microcline crystals may show an orientational arrangement, parallel to microcline {010} and {110}, as shown in figure 13. The occasional megacrysts also exhibit orientational control over their inclusions.

In comparing the modes of occurrence of the *quartz* in the Trancoso granite with those in the porphyritic granite it is noted that the quartz in the Trancoso rock is largely of quartz III type, with a little quartz I and IV, and sparingly some quartz V and VI. Bipyramids of original high quartz are scarce. They reach 0.25 mm in diameter and are found only in plagioclase (which in places has slightly corroded them). Since most quartz appears to have recrystallized, it is inferred that these bipyramids are armored relics protected by a coat of plagioclase. The main quartz is in irregular grains

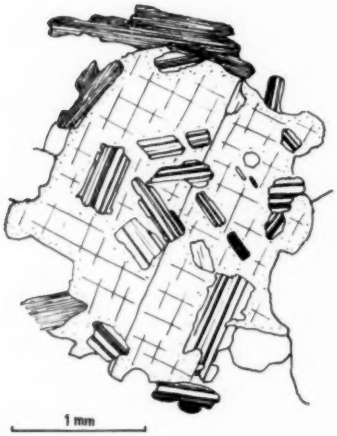

Fig. 13. Oriented plagioclase and biotite inclusions in new microcline. Trancoso granite.

which corrode biotite, plagioclase, and microcline, and show lobate boundaries (figs. 10, 11) in the granodioritic varieties. In these latter the quartz may also mold plagioclase without replacement. Drop-quartz is found in feldspar. Part of it may owe its isolated appearance to favorable section of lobed outgrowths of the main quartz grains. The lobes seem to be late metasomatic growths. The lobate quartz in figure 11a may have originated by coalescence of quartz drops metasomatically formed along an intergranular surface. In the more granitic specimens the quartz is irregular, still replacive in habit, but not lobate or but little so.

The quartz is generally undulatory. It is to be noted that cataclastic plagioclase may be associated with but slightly undulatory quartz. This points to recrystallization of the quartz after dislocation of plagioclase took place, with subsequent development of undulance.

Sillimanite has been observed in adamellitic specimens and also at the contact with the porphyritic granite northeast of Trancoso. It occurs chiefly in muscovite, rarely in quartz, and may replace biotite in the same way as in the porphyritic granite. In rock slices cut across the contact between the two granites trains of fibrolite may transect the contact, as does muscovite. This indicates that both minerals belong to the paragenesis of the younger granite.

Muscovite is latest among the rock components. It shows the same properties as in the porphyritic granite ($-2V = 35°$-$39°$, averaging $36.5°$; birefringence 0.033-0.035) and builds ragged plates (up to 4 mm in size), frequently grading into irregular vein-like forms penetrating between the older minerals. The cleavage of this interstitial muscovite lies at a large angle to the direction of the vein (fig. 14).

The history of the rock may be summarized as follows: crystallization of a hypidiomorphic granodiorite—slight deformation prior to or possibly contemporaneous with the intrusion of the porphyritic granite—extensive re-

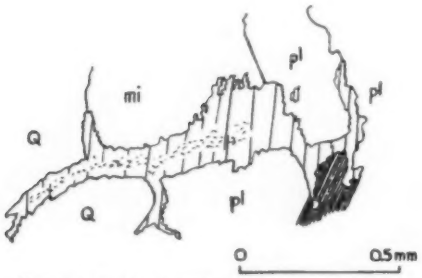

Fig. 14. (left). A train of fibrolite in late muscovite. Note interstitial attitude of muscovite. Trancoso granite.

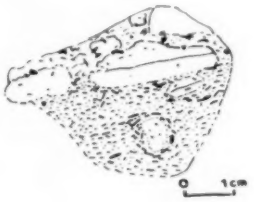

Fig. 15. (right). Contact between porphyritic granite and gneissose xenolith. Note parallelism of biotite flakes, wrapping around microcline megacrysts.

crystallization and metasomatism, resulting in albite-oligoclase, microcline, quartz, sillimanite, and muscovite.

CONTACT RELATIONS

Although the contact between the two granites is generally obscured by superficial deposits, an intrusive relationship was established. Where observed, the contact is sharp. West of Trancoso the porphyritic granite carries a few inclusions of older granite near the contact. Near Souto Maior, northeast of Trancoso, the porphyritic granite sends out veins into the Trancoso granite. Three veins were observed, varying in width from 2 to 4 meters and dipping into the porphyritic granite at angles of 30° to 50°. These appear to be infillings of marginal fractures in the granodiorite caused by the rising magma of the younger granite. That the porphyritic granite is the younger is also proved by the extensive modification the granodiorite has undergone.

XENOLITHS

Inclusions or xenoliths are extremely rare in the younger granite and absent in the Trancoso granite. Near contacts, the porphyritic granite may carry a few angular inclusions of the adjacent rock. About 1 km WNW of Trancoso a large xenolithic mass, perhaps a hundred meters across, lies between the two granites. The contact with Trancoso granite is not exposed. The contact with porphyritic granite is sharp and dips 55° ESE. Here many pegmatite veins occur, chiefly in the xenolith. This large inclusion consists of a gray speckled rock with thin biotite layers and quartzofeldspathic bands, with gneissosity striking 45°-50° east of north, and vertical dip.

The rock is a fine-grained biotite-rich gneiss, mainly consisting of granoblastic quartz and plagioclase and idioblastic biotite plates (in subparallel orientation). The accessories are formed by minute zircon and zircon-like crystals, with little apatite and no ore, distributed indifferently within quartz, feldspar, and mica. Abundant quartz forms both rounded and irregular grains (up to 1 mm); the latter embay the plagioclase, occasionally in lobate forms. Plagioclase consists of albite-oligoclase (An_{9-12}). Muscovite occurs in a few small flakes intergrown with biotite or irregularly surrounding and corroding

biotite. The absence of sillimanite is to be noted. In the center of the mass there is no microcline; this is developed only near the contact with the porphyritic granite. This plagioclase gneiss is probably of sedimentary origin.

Up to about 80 meters west of this large xenolith, the porphyritic granite contains occasionally a few small xenoliths of microclinized plagioclase gneiss, along with some much larger angular inclusions of Trancoso granite (up to 2 or 3 meters across). The gneiss inclusions are rounded and disk-shaped, in sizes up to 40 cm. Although the porphyritic granite contains abundant muscovite, these xenoliths are very poor in it. They show the same texture as the large gneiss mass but carry microcline in addition. The composition of the plagioclase is albite-oligoclase (An_{7-13}). The microcline is in irregular grains (up to 0.5 mm in size) which embay the plagioclase. The microcline shows undulance and coarse cross-hatch twinning, and occasionally exhibits a little vein perthite or, rarely, some film perthite. The quartz and the plagioclase are undulatory also. There is in addition a development of sporadic microcline megacrysts within the xenoliths (fig. 15). At the microcline stage of the surrounding porphyritic granite these xenoliths seem to have been pervaded by the late granitic fluids separating potash feldspar. Local accumulation may have formed megacrysts. The xenoliths were reduced to a plastic, part-fluid condition at this stage and consisted of a mash of biotite, plagioclase and possibly some quartz, permeated by late fluids. This is indicated by the biotite foliation wrapping around microcline megacrysts in the xenoliths and around megacrysts projecting from granite into xenolith (fig. 15). A relatively high degree of plasticity is indicated hereby, so that the mica flakes were pushed aside by the growing megacrysts. When microcline had separated, the xenoliths apparently became almost impervious to the fluids of the metasomatic stage in the granite, so that they contain but little muscovite.

THE PEGMATITES

Pegmatites are abundant around Trancoso. Their greatest concentration is found in the older granite along the two granites contact. Aplite has been observed only locally, in porphyritic granite northwest of Trancoso, in thin veins (1-5 cm) apparently on cross-joints, striking about N 55 with vertical inclination.

The pegmatites generally form irregular, pinching and swelling veins, lenses, and sheets, varying in width from a few centimeters up to 3 meters. They consist of coarse irregular quartz, potash feldspar (up to 20 cm in size) and muscovite (up to 6 cm). There is frequently also some biotite. The mica crystals often display six-sided outlines. The biotite is frequently surrounded and corroded by muscovite. Black tourmaline is occasionally met with, while pale green beryl has been noted in places (west of Trancoso). The beryl is euhedral against quartz and molded on potash feldspar. The quartz is generally milky. Small euhedral crystals of smoky quartz, lining cavities, occur here and there. The potash feldspar is in general euhedral. It is tabular on 010 and displays the forms {010}, {001}, {110} and {10$\bar{1}$}. It is corroded by quartz and muscovite and carries unoriented albite inclusions. The potash feldspar is perthitic (chiefly vein perthite grading into patch

perthite; a little film perthite is occasionally present). It consists in the main of untwinned microcline ($2V\alpha = 80°\text{-}85°$). Orthoclase has also been noted ($2V\alpha$ about 60°).

The boundaries of the pegmatites are vague, merging into the surrounding granite. Relatively sharp borders may occur against Trancoso granite. In this granite many of the pegmatites are in sheets with low dips toward the porphyritic granite. The interior of the Trancoso massif which is most granodioritic in composition appears to contain few or no pegmatites.

The pegmatites seem to have formed from the rest-fluids of the porphyritic granite.

MICROCLINIZATION OF TRANCOSO GRANODIORITE

The increase in microcline content in the Trancoso granite from the interior of the body towards the periphery (compare the locations of the modally analyzed specimens in figure 1) is not due to initial variability in composition. It is a secondary feature brought about by extensive development of new microcline near the porphyritic granite. The old microcline in the granodiorite, with smaller optic axial angle, may represent the original potash feldspar of the rock, most probably entirely recrystallized. It may be pseudomorphic after the original potash feldspar, just as the albite-oligoclase is considered to be pseudomorphic after more calcic primary plagioclase (see below). It is thought that the granodiorite is not a microclinized earlier trondhjemite.

The increase in microcline content toward the porphyritic granite points to an influx of potash feldspar material from the younger granite, which is confirmed by the replacive habit of the microcline. Concomitant with the increase in microcline, the old microcline disappears, recrystallizing as new microcline. The new microcline is thus in part metamorphic and in part introduced. The same relationship probably holds for the quartz. Although it would appear from the modal analyses that the new microcline developed wholly at the expense of plagioclase (and by recrystallization of old microcline), this is not entirely true. On considering the fabric it becomes clear that the original granodiorite, a hypidiomorphic biotite-plagioclase-potash feldspar-quartz rock, became microclinized through the introduction of material replacing both plagioclase and original quartz. The new microcline made room for itself by replacing granodiorite minerals, quartz not excluded. Afterward quartz crystallized, partly replacing both plagioclase and microcline. The granodiorite was solid when the younger granite was emplaced. This follows from the sharp contacts. Apparently the metasomatic, crystalloblastic process by which the new microcline developed led to a partly fluid state, since orientation of its tabular biotite and plagioclase inclusions may be observed.

The microclinization is considered to have been a metasomatic contact effect of the porphyritic granite. The microclinized granodiorite may be regarded as a regional reaction rim between the porphyritic granite and the granodiorite proper. The process seems to have been effected by an emanation of mobile substances into the predominantly solid, recrystallizing gran-

odiorite. The presence of late muscovite in large quantities indicates a high degree of fluxing by hydroxyl which possibly was instrumental in bringing about the metasomatic processes transforming the original granodiorite. The interstitial attitude of the new minerals in their early stages strongly suggests that the influx took place along intergranular surfaces, in fluid form. The deformation which is locally evidenced by cataclastic plagioclase probably also effected as a less intense but more extensive process of dislocation a general loosening of the grains of granodiorite minerals. This may have facilitated imbibition with fluids soaking in from the porphyritic granite magma along the loosened intergranular spaces of the slightly dislocated granodiorite fabric. Thus the impregnation of the rock gave rise to new microcline, quartz, and muscovite. Though this process, as a kind of palingenesis, may have started before the younger granite magma had completed intrusion, its activity must have taken place largely after emplacement of the younger granite since its mineral facies are mainly those of the later magmatic and the metasomatic stages of the porphyritic granite. The development of microcline is probably an early phase of the process. The replacement of older minerals seems to have given rise, at least locally, to pockets of fluid in which inclusions were oriented by the separating microcline. Larger pockets of fluid permitted the formation of megacrysts, also with oriented inclusions. This fluidization, probably due to fluxing, correlates this stage with the magmatic stage of the porphyritic granite. A process in which a new mineral begins crystallization by blastesis, to continue separation in and from a partly or entirely fluid environment, might perhaps be designated by the term *ultrablastesis*. It is not ultrametamorphic, because new material was metasomatically introduced. Just as the stage of microcline formation in the porphyritic granite was reached under falling temperature in a consolidating magma, so the new microcline in the granodiorite developed under rising temperature, leading to partial fluidization of the rock. Afterward, when the temperature was declining again, quartz and muscovite crystallized. The optic axial angle of the granodiorite microcline is smaller than in the microcline of the porphyritic granite. This may perhaps point to its having formed at somewhat higher temperatures. The persistence of smaller axial angles may indicate incomplete adjustment to the conditions of later stages.

DECALCIFICATION OF PLAGIOCLASE

The uniform albite-oligoclase composition of the plagioclase in all plagioclase-bearing rocks in the area appears to be due to an extensive process of decalcification.

The large plagioclase crystals in the Portuguese porphyritic granites I have examined (such as the Cete granite) frequently display normal zoning with cores as calcic as andesine. The texture of the porphyritic granite around Trancoso closely resembles the texture of the other porphyritic granites. It is therefore inferred that the primary plagioclase in this granite also exhibited a composition more calcic than albite-oligoclase, but was afterwards decalcified.

Persistent zonation in the plagioclase of igneous rocks, denoting internal disequilibrium, may be considered not to be due to lack of reactivity but to lack of time to effect readjustment in response to changing conditions. In the granite in question the temperature may be assumed to have decreased sufficiently slowly to allow the reaction from more calcic to more sodic plagioclase to proceed to completion. If the rock was kept for a considerable period at the pt-conditions of the albite-oligoclase stage in its crystallization history, the primary plagioclase could have been entirely decalcified to homogeneous albite-oligoclase pseudomorphs. This may conceivably be attributed to a slow rate of cooling. The porphyritic granite around Trancoso is situated in the center of an enormous batholith. Such a pseudomorphic process would act in the solid, most probably by ionic diffusion in the plagioclase framework.

In view of this, it also seems likely that the plagioclases of the granodiorite and of the gneiss xenoliths, whatever their original composition, were metasomatically changed into albite-oligoclase, since they may be considered to have been raised to the same mineral facies as obtaining in the huge surrounding mass of porphyritic granite. The restricted range of anorthite content in the three types of rocks is too similar to be fortuitous. In the granodiorite and the gneiss inclusions, albite-oligoclase formation is therefore considered a contact effect of the younger porphyritic granite.

Where the plagioclase is in contact with microcline it has albitic outer zones. The transition is abrupt in the granodiorite, gradual in the other granites. This seems to be due to a decalcifying process of essentially the same nature as the homogenization to albite-oligoclase, but not run to completion since it is zonal. It is curious that these albite rims occur only in contact with microcline. Since they are absent against adjacent quartz III which corrodes microcline, they were probably formed posterior to the separation of potash feldspar. In the porphyritic granite the formation of the albite rims may be assigned to the metasomatic stage. Albite is not a magmatic mineral in these rocks (Schermerhorn, 1956, where it has been shown that the magmatic passed into the metasomatic stage approximately at the transition from the conditions of the hornblende gabbro facies to those of the albite-epidote amphibolite facies). It is thus indicated that after microcline had formed, albite-oligoclase was no longer the stable phase of plagioclase in the metasomatic stage and was supplanted by albite. The abrupt narrow zones of decalcification in the granodiorite indicate that here the readjustment possibilities were much more restricted than in the porphyritic granite and in the more microclinized varieties of Trancoso granite.

Concerning decalcification, it may be remarked that the plagioclase in the rocks of the area under consideration carries few or no zoisite-epidote minerals. It is concluded that the decalcifying process consisted not only of the release of Ca and Al (as commonly in the decomposition of calcic plagioclase to sodic plagioclase and zoisite-epidote) but also of their removal. We may infer that the amount of Ca and Al set free during decalcification passed into the rest-fluids which yielded Na and Si to replace Ca and Al. Part of the Al released may immediately have gone toward formation of sericite and other alteration products of plagioclase. Furthermore, it seems logical to

correlate the sphene replacing ilmenite and biotite with the decalcification of plagioclase. It may be conceived that owing to the calcium content of the rest-fluids, sphene was formed from Ti (or sagenite) released by biotite or derived from ilmenite. In the Cete granite (Schermerhorn, 1956) on the other hand, exhibiting a much lower degree of decalcification, there is but little late sphene. The aluminum in the rest-fluids went toward crystallization of sillimanite and late muscovite.

METASOMATIC SILLIMANITE AND MUSCOVITE

Sillimanite is rarely encountered in igneous rocks, where it is currently considered due to contamination by pelitic material. As a late mineral it is occasionally noted in granites and pegmatites. Hills (1938) lists a few occurrences of sillimanite-bearing igneous rocks assumed to have crystallized from uncontaminated magmas. Read (1931, p. 162-163) describes sillimanite-bearing muscovite granites in Sutherland. Read considers that the excess alumina needed to form sillimanite was contained in the granitic fluids and was not the result of assimilation: "The sillimanite and the replacing or late muscovite . . . are genetically connected, and both most likely arise by the break-up of aluminum halides, alkali aluminates, and similar compounds contained in the residual portions of the granitic magma." Watson (1948) further investigated sillimanite occurrences in the Central Sutherland injection complex, both in country rock and in pegmatitic and aplitic veins. This sillimanite is convincingly shown to have formed under the influence of pegmatitic solutions, at a late stage, as a metasomatic, not a metamorphic mineral.

The mode of occurrence of sillimanite in the two granites of Trancoso is indicative of late origin. The occasional vein-like habit of the muscovite with which it is associated and the interstitial attitude of sillimanite trains suggest permeation of fluids along intergranular spaces. It is thought that the muscovite was largely introduced from the porphyritic granite, although part of it may have been derived from an initial muscovite content of the granodiorite, in veins following the boundaries of feldspar and quartz grains, and starting replacement from between the older crystals. The muscovite grew at a high angle to the intergranular spaces as evidenced by cleavage attitudes. The sillimanite seems to have formed prior to muscovite. Possibly it is only locally present owing to local alumina excess or rather potash shortage in the late fluids. Sillimanite and muscovite are both late, metasomatic minerals that crystallized from the residual fluids given off by the porphyritic granite. It is most unlikely that the sillimanite was derived from assimilated country rock, since the sparse xenoliths do not carry alumino-silicates. Nor are there any indications of contamination in the granites. The relatively large quantity of secondary muscovite, on the other hand, points to the existence of peralumious rest-fluids of magmatic derivation. Relative alumina enrichment, to separate sillimanite before muscovite formed, might have been caused by local disturbances of the K-Al balance. The alumina that went to form sillimanite in the Trancoso granite may have formed as a consequence of decalcification of plagioclase. This sillimanite has nothing to do with metamorphic sillimanite as an index mineral for high grade schists or horn-

felses. In the crystallization history of the Trancoso rocks it is posterior to quartz III. The sillimanite here was therefore formed in a late stage in their evolution, as a metasomatic mineral.

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A RECENT VOLCANIC MUDFLOW OF EXCEPTIONAL DIMENSIONS FROM MT. RAINIER, WASHINGTON*

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ABSTRACT. A Recent mudflow extends from the northeast side of Mt. Rainier 45 miles down the White River valley to the west front of the Cascade Range and spreads out on the Puget Sound lowland east of Tacoma as a lobe 20 miles long and 3 to 10 miles wide. The mudflow originally was named Osceola till and described as the deposit of a piedmont glacier originating in the Cascade Range during the Vashon glaciation.

The mudflow deposit ranges in thickness from a few feet to 350 feet and typically is unsorted and unstratified. It is characterized by a purplish-gray color, by abundant andesite lava fragments from Mt. Rainier, and by wood fragments ranging from logs 10 feet long to very fine fibrous particles. In the lowland the mudflow lobe is bisected by the trench of the White River; margins of the flow are digitate, with narrow lobes extending as much as 4 miles beyond the main body. The upper surface, although appearing flat over broad areas, has a northwesterly slope of 25 to 30 feet per mile.

In the lowland the mudflow overlies a brown podzolic soil profile developed on Vashon glacial drift; on Mt. Rainier the mudflow is postdated by at least three eruptions of pyroclastic material and by at least two advances of Emmons Glacier. Carbon 14 dating of two wood samples provides an approximate age of 4800 years for the mudflow.

The mudflow is thought to have originated as a result of lateral ejections of clayey material from the northeast side of Mt. Rainier. The name Osceola mudflow is here applied to the deposits included in the Osceola till of former usage.

INTRODUCTION

An unsorted, till-like deposit of boulders and smaller stones in a clayey and silty sand matrix underlies some 50 square miles of the Puget Sound lowland about 20 miles east of Tacoma, Washington (fig. 1). The deposit extends from the lowland up the valley of the White River in the Cascade Range to the northeast side of Mt. Rainier.

This deposit was first described by Willis (1898, p. 143) as "a sheet of till which covers the plateau between Green and White rivers and extends southwest beyond White river about the head of Fennel Creek." He proposed (p. 130) that "this till be referred to as the Osceola till, as that hamlet lies at its widest expanse," and suggested that the till was deposited by a lobe of a glacier originating in the Cascade Range to the southeast. Willis described the till (p. 143) as a "fine silt with numerous angular stones. The silt is a glacial meal such as White river carries at the present time. The stones are prevailingly, if not almost exclusively, fragments of Tertiary or later volcanic rocks."

Willis recognized the exclusively Cascade provenance of the deposit which differentiates it from the most recent glacial drift of northern derivation, to which he gave the name Vashon. Willis believed his Osceola till underlay the Vashon drift, and that the two formations belonged to essentially one episode of glaciation. He concluded that the Osceola was deposited by a piedmont glacier which moved northwestward from the Cascade Range and became confluent with the Vashon ice sheet which advanced into the area from the north.

The only subsequent critical evaluation of the so-called Osceola till is that of Bretz (1913, p. 23, 223), who accepted Willis' interpretation of the origin of the deposit in the lowland area but disagreed with the southwesterly

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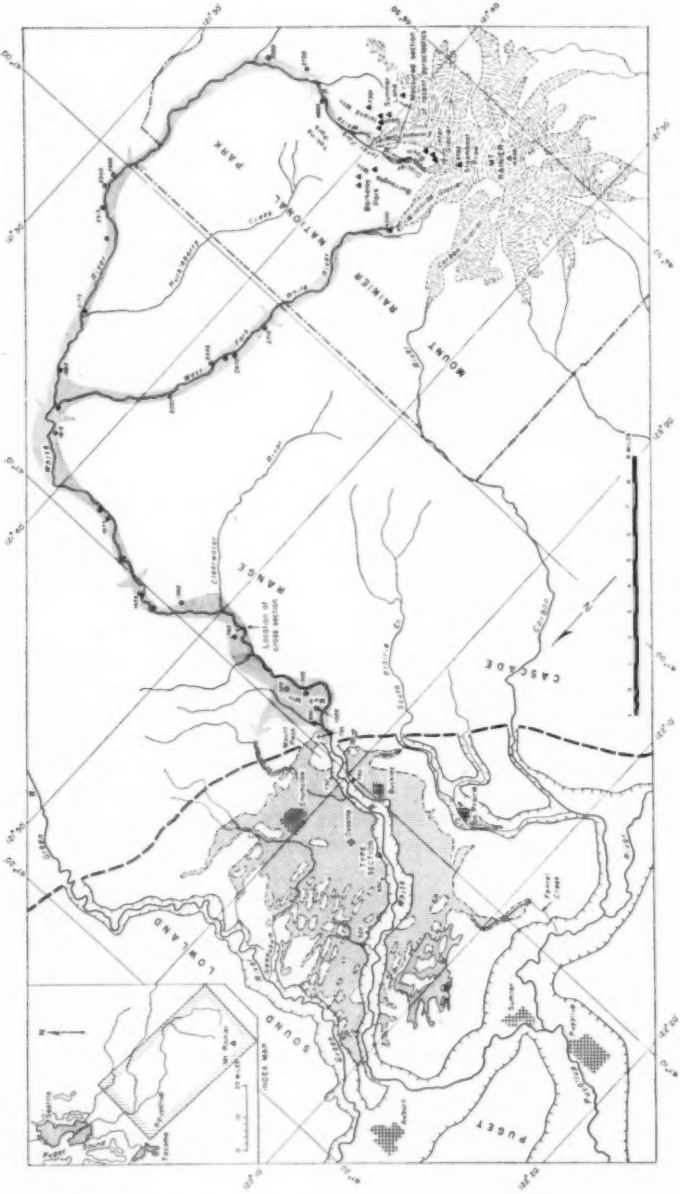

PARK
NATIONAL
RAINIER
MOUNT
RANGE
CASCADE
LOWLAND
SOUND
PUGET

Fig. 1

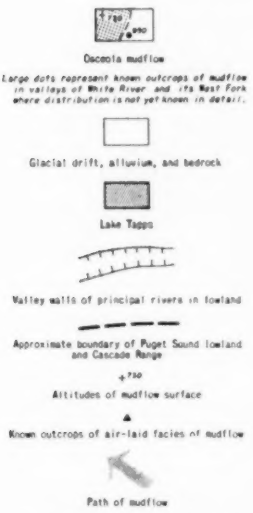

Osceola mudflow
Large dots represent known outcrops of mudflow in valleys of White River and its West Fork where distribution is not yet known in detail.
Glacial drift, alluvium, and bedrock
Lake Tapps
Valley walls of principal rivers in lowland
Approximate boundary of Puget Sound lowland and Cascade Range
+730
Altitudes of mudflow surface
Known outcrops of air-laid facies of mudflow
Path of mudflow

Fig. 1. Map showing distribution and path of Osceola mudflow. Areal distribution along north edge of mudflow taken from soils map of King County (Paulson and others, 1952) and locally revised by D. R. Mullineaux, U. S. Geological Survey and the writers. Base map from U. S. Geological Survey maps of Tacoma, Lake Tapps, Cedar Lake, and Mt. Rainier quadrangles.

extent of the till as it is shown in the Tacoma folio (Willis and Smith, 1899). Subsequent to Willis' and Bretz' work, the importance of Cascade valley glaciers as contributors of ice to the Puget Sound lowland ice sheet during the Vashon glaciation has been seriously questioned. Mackin (1941, p. 470) pointed out that at the maximum stand of the Vashon glacier, most valley glaciers on the west side of the Cascades were small and not confluent with Vashon ice.

Recent geologic mapping by the writers has shown that the Osceola till of Willis is considerably younger than the Vashon glacial drift, and that the deposit required some transporting medium of which essentially no part was above the present surface of the deposit. This requirement makes deposition by glacial ice highly improbable and suggests an origin by flowage. The distribution and character of the deposit indicates that it is a mudflow of volcanic origin that originated at Mt. Rainier.

The investigation of this deposit is being undertaken in conjunction with detailed geologic mapping in the Puget Sound lowland, which is yet incomplete. This paper is presented as a progress report inasmuch as many of the salient facts regarding distribution, topography, stratigraphic relations, and age in years of the mudflow are now established.

DESCRIPTION OF DEPOSIT

Internal appearance and composition.—The deposit is an unsorted mass of subangular to subrounded pebble, cobble, and boulder-sized fragments as large as 8 feet in diameter in a plastic clayey and silty sand matrix. The only semblance of sorting in the deposit is a marked decrease in both number and size of stones from bottom to top (pl. 1-A). This graded size distribution is seen at virtually every locality in the lowland where both bottom and top are

exposed, and evidently is characteristic of the deposit. Interbedded sand and gravel have not been seen, and nowhere is there evidence that the mudflow consists of more than a single unit.

The stones are principally andesite and basalt derived from the Keechelus andesitic series of Oligocene and Miocene age, which is widely distributed in the central Cascades of Washington, and of hypersthene and pyroxene andesite from lava flows and pyroclastic deposits of Mt. Rainier. A small percentage of the stones are derived from the Snoqualmie granodiorite of Oligocene(?) age, which crops out in the valley of the White River near its headwaters. The proportion of the several rock types, as determined by averaging pebble counts from eight localities in the lowland, is given below.

	Percent
1. Keechelus andesitic series	41
2. Andesite from Mt. Rainier	35
3. Rocks derived from either 1 or 2 above	10
4. Snoqualmie granodiorite	2
5. Other	12

A sample of the deposit from Glacier Basin on Mt. Rainier (fig. 1) was examined in the laboratories of the U. S. Geological Survey for mineralogical composition (table 1). A characteristic feature of the mudflow at this locality is the presence of fresh, euhedral hypersthene, which occurs in bundles of crystals as well as individuals. Some hypersthene grains have a skin of volcanic glass adhering to them, whereas others are embedded in a fine, felted matrix. The feldspars are fresh and many show zoning.

The unoxidized color of the mudflow when moist ranges from grayish purple to medium or light gray with a faint purplish cast; it oxidizes to mottled yellow or yellowish brown.

Cavities ranging in size from less than a millimeter to several millimeters are characteristic of the mudflow deposit. Exceptionally, cavities a foot or more in diameter and several feet in length are present near the base of the mudflow; these commonly are lined or partly filled with iron oxides. Some of the cavities very likely are molds of vegetal matter now rotted out, others probably represent air entrapped in the mudflow, and still others may be molds of ice fragments derived from glaciers on Mt. Rainier.

Another characteristic component of the deposit is wood debris. This ranges from logs several feet in diameter and at least 10 feet in length to wood chips and splinters, pine cones, and macerated vegetal matter. No stumps or overridden trees in the position of growth have yet been found. Several logs seen in lowland exposures of the mudflow are charred, but these are exceptional; most of the wood either is fresh or shows only normal decay.

Distribution and topography.—Willis (1898, p. 143) stated that Osceola till crops out between Green and White Rivers and about the head of Fennel Creek, but the Tacoma folio (Willis and Smith, 1899) also shows the so-called till north of the Green River, west of the Puyallup River south of Orting, and mantling much of the Cascade foothill area in the southeastern corner of the Tacoma quadrangle. The areas north of the Green and west of the Puyallup

TABLE 1

Grain-size Distribution and Mineralogical Composition of Mudflow Sample from Glacier Basin

Grain size mm	Percent	H.R. percent of total	Minerals and rocks present
2	24.8	n.d.	Volcanic rock fragments
1	4.8	n.d.	Rock fragments
0.5	6.1	n.d.	Rock fragments, pyrite, volcanic glass
0.25	8.1	0.5	H.R. Hypersthene; green pyroxene (abundant); rock matrix, partly glassy; pyrite; magnetite. L.F. Feldspar (labradorite-bytownite); rock matrix, partly felted and partly glassy (abundant)
0.105	9.8	0.9	H.R. Hypersthene (abundant), pyrite, ilmenite, olivine, magnetite. L.F. Glassy and crypto-crystalline matrix, labradorite-bytownite, volcanic glass
0.050	6.6	0.8	H.R. Hypersthene (very abundant), pyrite, olivine, magnetite. L.F. Feldspars, labradorite-bytownite (abundant), rock matrix grains, volcanic glass
Clay and silt	39.6	n.d.	Silt—Cristobalite and feldspar (abundant), montmorillonite, quartz Clay—Montmorillonite (abundant), kaolin mineral, hydrous mica, cristobalite, feldspar, quartz
Total	99.8		

H.R.—heavy residue
L.F.—light fraction
n.d.—not determined

Rivers are now known to be underlain by Vashon glacial drift; the higher parts of the Cascade foothills in the Tacoma quadrangle are underlain by a pre-Vashon till of Cascade derivation. It is likely that Willis and Smith interpreted this earlier drift as the foothill equivalent of their typical Osceola till in the lowland to the northwest.

In the Puget Sound lowland the deposit forms a broad, digitate lobe bisected by the valley of the White River (fig. 1). The eastern edge of the deposit terminates abruptly against the foothills of the Cascades except at the valley of White River where the deposit extends into the mountains nearly to the present terminuses of Emmons, Winthrop, and Inter Glaciers on Mt. Rainier. A hasty reconnaissance of the other major valleys heading on Mt. Rainier has not disclosed the existence of the material in any of these valleys.

In the lowland, many hills protrude through the surface of the deposit; some of these, such as Mount Peak (fig. 1), are drift-veneered bedrock knobs; others are till drumlins and sand and gravel deposits of Vashon age. Long lobes of the deposit partly fill pre-existing topographic depressions and extend as much as 4 miles beyond the margins of the main body. The best development of these is found north of Enumclaw and in the valley of Fennel Creek.

The surface of the deposit is remarkably flat over broad areas (pl. 1-B), although locally there are low swells and depressions. Most of these probably represent differential compaction over an irregular underlying surface. The deposit terminates sharply against hills and low ridges, but locally it occurs

PLATE 1

A. Exposure of mudflow resting on Vashon stratified drift at Mud Mountain. Note the general decrease upward in number and size of stones and their subangular to subrounded shape. Upper part of stratified drift contains a brown podzolic soil profile. Pick gives scale.

B. Aerial view of mudflow surface northwestward from above White River near Buckley. Forested areas in middle distance are underlain by Vashon drift.

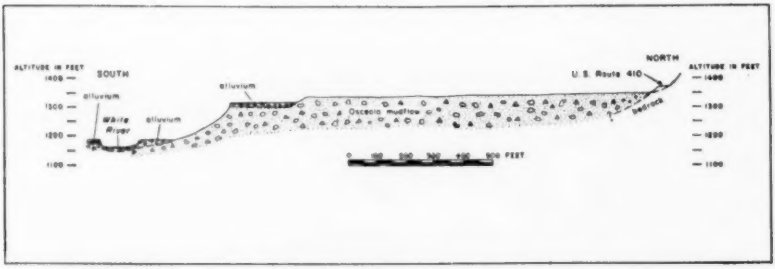


Fig. 2. Profile and section from White River floodplain north to U. S. Route 410 east of Mud Mountain (fig. 1). No vertical exaggeration.

above, below, and on low topographic scarps, which almost without exception face northwestward, the general direction of flowage.

The surface gradient of the deposit is 25 to 30 feet per mile toward the northwest, the downvalley direction of the White River. The gradient of individual lobes generally is in a direction away from the main body of the deposit.

Stream dissection of the surface of the deposit has been very minor, due chiefly to the low surface relief and slope and to the youthfulness of the deposit. In the lowland, gullies along the White River rarely have cut headward more than a few hundred yards, and most have a drainage area of a few tens of acres or less. The only stream that drains a large area is Newaukum Creek.

In the valleys of White River and West Fork in the Cascade Range the deposit is found in fill-terrace remnants (fig. 2); some of these fills retain depositional upper surfaces (pl. 3), whereas others have been modified by subsequent stream erosion and deposition. The gradient of the terrace remnants increases upstream from 45 feet per mile near Mud Mountain Dam to more than 400 feet per mile in Inter Fork valley, and roughly parallels the present gradient of White River. In the valley of West Fork the deposit was traced upstream to less than a mile from the present terminus of Winthrop Glacier. Owing to the inaccesibility of the area, however, very little is yet known of the distribution of the mudflow in this drainage basin. In Glacier Basin (fig. 1), a tributary valley of White River that heads in Inter Glacier, the deposit occurs as a dissected blanket as much as 100 feet thick over the floor and lower slopes of the basin. On the ridges surrounding Glacier Basin are many small, scattered outcrops of the deposit; the highest of these is at the top of Steamboat Prow. Material resembling the deposit in color and stratigraphic position also has been found at many localities on the summit and lower slopes of Goat Island and Burroughs Mountains, and in Berkeley Park, Summer Land, and Yakima Park (fig. 1).

Thickness and volume.—The deposit attains a maximum exposed thickness in the lowland along the north walls of the White River valley downstream from Buckley in secs. 29, 33, and 34, T. 20 N., R. 6 E. There, in nearly vertical bluffs, it overlies Vashon glacial drift and ranges in thickness from

PLATE 2

A. View east down Inter Fork valley to White River valley from Burroughs Mountain. Forested spur separating the two valleys is in part a pre-mudflow lateral moraine of Emmons Glacier, which constricted the mouth of Inter Fork valley. Light-colored exposure in center of photograph is shown in plate 3.

B. View of Mt. Rainier from Yakima Park showing moraines of Emmons Glacier and mudflow terrace in Inter Fork valley.

25 to 75 feet. Farther downstream the deposit in places pinches out against knobs of glacial drift too high to have been inundated. At distances of more than a mile from the White River valley the deposit generally is less than 20 feet thick.

An inverse relationship between valley width and thickness of the deposit occurs at two localities in the Cascade Range segment of the White River valley. At Mud Mountain the river has cut a narrow gorge nearly 350 feet deep into bedrock and glacial drift. This gorge is now partly occupied by Mud Mountain Dam. Immediately upstream from this gorge the deposit is 350 feet thick, although still farther upstream the deposit rarely is more than 100 feet thick. It is inferred that the constriction of the valley at the gorge acted as a temporary and partial obstruction to the mudflow, so that some of the material accumulated as a "backwater" upstream. This "backwater" eventually rose to the level of a wide, natural spillway (the top of Mud Mountain); passage of mud over this spillway prevented the accumulation of a greater thickness upstream. At the valley mouth of Inter Fork, a lateral moraine of Emmons Glacier (pl. 2-A) has partly blocked the valley. Upstream from this constriction the deposit is 250 feet thick (pl. 3); although both upvalley and downvalley it generally is again less than 100 feet thick.

Adequate data are not yet available concerning the average thickness of the deposit, but if an average thickness of 20 feet is assumed in the lowland, the reconstructed volume there alone is more than 1 billion cubic yards. In addition, an undetermined volume exists in the 40-mile segment of the White River valley and the 17-mile segment of West Fork valley between the lowland and Mt. Rainier. It is believed that the total volume was at least 1.5 billion cubic yards, but a reliable figure awaits more detailed investigation.

Stratigraphic relations and age.—Contrary to the inference of Willis, the deposit overlies Vashon glacial drift in the Puget Sound lowland. At many localities the upper part of the glacial drift is marked by a moderately well-developed brown podzolic soil profile (pl. 1-A), which has been buried by the deposit. Locally the two deposits are separated by a layer several inches thick of highly compressed vegetal matter. The podzolic soil profile represents a period of soil formation between the deposition of the Vashon glacial drift and the occurrence of the deposit, and so it can be inferred that the two deposits are separated by an appreciable time interval.

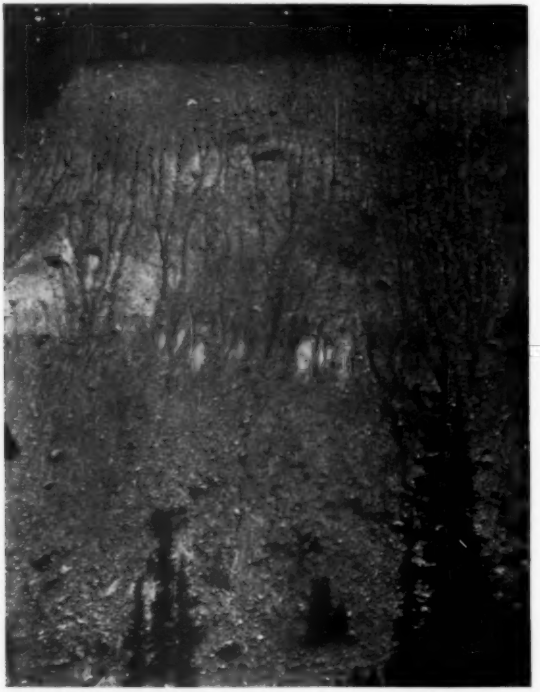
In the lowland the soil on the deposit is highly organic and is dark grayish brown to nearly black over broad areas of low relief. Below this surface layer of about 8 inches is a zone, about 6 inches thick, of oxidized material which is lighter in color and lower in content of organic matter. This grades downward into gray and yellowish-gray material mottled with brown iron oxides. The iron oxides, and probably some colloidal silica, act as a cement in the deposit, locally causing the upper 10 feet or so to be well indurated. In several places the deposit is overlain by peat and associated lacustrine or swamp deposits and stream alluvium.

On the northeastern side of Mt. Rainier the deposit is overlain by as much as 6 feet of ash and cinders, in which at least three pyroclastic eruptions are represented. Matthes (1914, p. 9) stated that Mt. Rainier "is known to

have been active at intervals during the last century, and actual record exists of slight eruptions in 1843, 1854, 1858, and 1870. Indian legends mention a great cataclysmic outbreak at an earlier period." Matthes inferred further that the date of production of the summit crater probably falls well within historic time.

A typical section of the Recent pyroclastics is given below. The section was measured in a pit dug at the crest of a lateral moraine of Emmons Glacier at the locality indicated on figure 1, where the slope of the moraine crestline is 14°.

PLATE 3



Exposure of mudflow in Inter Fork valley just upstream from constriction in valley caused by lateral moraine (see pl. 2-A). Mudflow is about 250 feet thick here. Note flat terrace surface preserved on top of mudflow.

	Inches
8. Fine-sand-sized ash, yellowish brown, weak podzolic soil profile at top	0.5-5
7. Cinders, pumiceous, brown and light gray, pea-sized to 3 inches in diameter	13.
6. Medium- to fine-sand-sized ash, dark grayish brown, with disseminated coarse-sand-sized ash	26.

5. Sandy and silty clay, highly plastic, yellowish at base, grades up to yellowish brown at top (air-laid mudflow equivalent)		9.5
4. Medium- to fine-sand-sized ash, dark yellowish brown to olive brown		5.
3. Silt-sized ash, yellow		1.5
2. Medium-sand-sized ash, yellowish brown		0-12
1. Till	More than	12.

On the basis of stratigraphic relationship to the Vashon glacial drift and a podzolic soil formed on it, and on the assumption that this post-Vashon soil was formed during the Thermal Maximum, the writers (Crandell and Waldron, 1954) previously inferred that the deposit is less than 5000 years old. Two samples of wood obtained from logs within the deposit were submitted to the Lamont Laboratory for C^{14} age determinations, and the following results were reported (W. G. Schlecht, U. S. Geological Survey, written communication, Oct. 7, 1954):

Lamont No.	U.S.G.S. No.	Approximate age (years)
223A	9CC-100 (clean wood)	4800±300
223B	1LC-100 (decayed wood)	4950±300

The deposit thus appears to have occurred during the Thermal Maximum, which extended from about 6000 to 3000 years ago (Flint, 1953, p. 174). Hansen (1947, p. 113) stated that pollen studies indicate a climatic trend toward cooler and moister conditions in the Pacific Northwest since the Thermal Maximum; these conditions are correlative with the little ice-age of Matthes (1942, p. 377). The fresh, conspicuous lateral and terminal moraines of Emmons Glacier (pl. 2-A and B) mark the second of two glacial advances since the Thermal Maximum and probably were formed within the past 200 years.

ORIGIN

The distribution, lithology, and age of the deposit indicate that it was deposited as a mudflow of volcanic origin rather than as glacial drift from a Cascade piedmont glacier. The criteria believed to be most important in determining this particular deposit as of nonglacial origin are summarized as follows:

(1) The presence of graded size-distribution in the deposit. It is conceivable that till locally might have apparent graded size-distribution; owing to the manner of deposition, however, such vertical size-distribution could hardly be as widespread or as distinct in till.

(2) The presence of the deposit in depositional terrace remnants in the valley of the White River. Till deposited by a valley glacier would be found as a more or less continuous mantle extending up the valley sides to the maximum height attained by the glacier.

(3) A lack of any morainal topography on the deposit. The writers are familiar with till plains of very low relief in the midwestern United States and recognize that lack of local relief in the lowland is not an independent

criterion. It is believed, however, that a valley and piedmont glacier hardly could have reached the extent indicated without building conspicuous lateral and terminal moraines somewhere along its periphery.

(4) Digitate distribution of the deposit.

(5) Evidence of a "backwater" resulting from restriction of flowage at Mud Mountain gorge and in the valley of Inter Fork.

(6) The fact that the deposit terminates sharply against hills and low ridges instead of overriding or destroying them as would be expected of a valley and piedmont glacier occupying the same area.

(7) Lack of any outwash sand and gravel associated with the deposit, as would be expected of drift deposited by a valley and piedmont glacier.

(8) Age of the deposit, dating it as contemporaneous with the Thermal Maximum, a time characterized by a dry, warm climate and a general retreat or disappearance of glaciers.

Several features of the mudflow indicate a volcanic origin for the deposit: the presence of material similar to the mudflow on the summits of Goat Island and Burroughs Mountains, and at Yakima Park, Summer Land, and Berkeley Park, where it could not have been transported by a mudflow and must therefore represent an air-laid facies of the deposit; the occurrence in Glacier Basin of fresh, euhedral hypersthene in bundles and individual crystals; and the apparent restriction of the mudflow to the valleys of the White River and West and Inter Forks, where the material extends from near the present terminuses of the Emmons, Winthrop, and Inter Glaciers down the length of the valleys. All these features suggest an eruptive source for the material from the northeast side of Mt. Rainier, but neither the exact source nor the mode of eruption that gave birth to the mudflow is known.

From the distribution of the material, the most likely source appears to be from a vent or vents lying somewhere between the summit of Mt. Rainier and Steamboat Prow, now buried beneath the extensive mass of ice that forms the upper reaches of Emmons and Winthrop Glaciers. The very existence of this broad expanse of ice suggests the possibility that the ice occupies a depression that resulted from the explosive destruction of part of the volcano, or from eruptive action followed by partial collapse of this area. Either of these two modes of eruption could have provided sufficient material for the mudflow.

Some of the features of the deposit suggest that it might be of peléan cloud origin, such as the evidence of high mobility, general restriction to topographic depressions, nearly flat upper surface over buried irregular surfaces, and apparent lack of sorting. However, the general absence of charring of included vegetal matter, the absence of pumice, and presence of graded size-distribution indicate that the deposit cannot be wholly the result of a peléan cloud or clouds. It is possible that the mudflow originated as a series of peléan cloudlike eruptions, but the mode of travel downvalley was a mudflow rather than as a mass of rock material in a gaseous suspension.

The clayey component of the mudflow very likely was derived from hydrothermally altered older rocks prior to eruption rather than from subaerial alteration of volcanic ejecta during and subsequent to the eruption. Rock

fragments in the mudflow very likely were derived from the upper slopes of Mt. Rainier, from alluvium on the valley floors, and from debris in and upon the glaciers. The sources of moisture may have been juvenile steam associated with the eruption, melting of glacier ice by steam (Kjartansson, 1951, p. 42-43), or rainfall.

REDEFINITION OF OSCEOLA TILL OF WILLIS

The Osceola till of Willis (1898) is here redefined as the Osceola mudflow. This mudflow is of Recent (Thermal Maximum) age and is restricted in geographic distribution to Mt. Rainier and the adjoining part of the Puget Sound lowland in the vicinity of White River.

A typical section of the Osceola mudflow is in a gully in the northeast valley wall of the White River about 3.5 miles west of Enumclaw, Wash. The gully is in the NE¼SE¼ sec. 29, T. 20 N., R. 6 E., and is about 0.2 mile west of the intersection of Wilson Road and Barker Road. A description of the typical section is given below.

	Feet	Inches
2. Osceola mudflow. Basal 15 feet consists of pebbles, cobbles, and boulders as large as 4 feet in diameter in a matrix of purplish-gray clayey and silty sand. Boulders and cobbles are mostly subangular to subrounded. Cavities ranging in diameter from less than 1 mm to 5 mm are abundant in matrix. Secondary iron oxides coat rock fragments and are disseminated throughout the matrix in streaks and nodular masses. Above the basal 15 feet, mudflow is cut by near-vertical joints, along and adjacent to which secondary iron oxides have impregnated and cemented the material to form resistant ribs 1 to 8 inches thick. About 25 feet above base of mudflow, rock fragments are predominantly of pebble and cobble sizes; this material grades upward into mudflow in which cobbles are rare, and in the uppermost 6 feet, the mudflow consists chiefly of sand and granule sizes with a few scattered pebbles. Wood fragments are scattered throughout section. No stratification can be seen, and no sorting occurs other than that described above. The base of the mudflow lies in a shallow depression in the underlying sand and gravel; the top of the deposit is the unmodified depositional surface of the mudflow, and forms the present ground surface.	75	
1. Stratified drift of Vashon age, in the upper part of which is a soil profile		
d. Humified silt and sand, black		0-2
c. Clayey and silty sand	1	8
b. Sand, medium-grained	2	4
a. Pebble and cobble gravel. More than	1	
More than	5	

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THE JURASSIC REEFS ON THE YONNE (SOUTHEASTERN PARIS BASIN)

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ABSTRACT. The Jurassic reefs on the Yonne are an example of a common type of reef formed in the geological past. They developed in an epicontinental sea during an atectonic and non-glacial period. They differ both from recent glacially controlled reefs and from reefs formed in mobile tectonic belts because they are much thinner. Moreover, the surrounding organic detrital limestone, oölite and fine-grained limestone are not part of a "reef-rock complex" with the reefs themselves, but formed independently.

INTRODUCTION

Reefs have occupied a prominent position in geologic research in recent decades. Recent reefs have been studied in various tropical regions, whereas fossil reefs have attracted the attention of both the surface and the subsurface geologist. The modes governing reef formation are so diversified, however, that we feel that even now not all important types of reefs have been described. To fill up part of this gap, a description of the Jurassic reefs on the Yonne river in the southeastern Paris Basin is presented.

The evolution of recent reefs has always been governed in a larger or smaller degree by glacial control. Moreover, tectonic movements, resulting from post-orogenetic activity of the Alpine paroxysm, have combined with glacial control to give the history of the present-day reefs a very lively character. In regard to fossil reefs, attention has centered predominantly on reefs in tectonically active areas. Here reefs have grown to a considerable thickness on subsiding basements, forming important reservoir rocks for the oil industry. Although several types of reef formation, either in the present-day reefs or in fossil reefs, have consequently become known in detail, their evolution was influenced mainly by climatic or orogenic control.

Long periods of the Earth's history are, however, characterized by being non-glacial and atectonic (Rutten, 1953). Shelf seas developing during these periods were predominantly epeirogenetic, epicontinental, and shallow, and were bordered by a virtually peneplained hinterland. Subsidence was slight and interrupted, being compensated for by minor uplifts of the hinterland. These seas were warm, with tropical conditions reigning to high degrees of latitude. Moreover, no abrupt cyclic temperature changes occurred, nor were there eustatic cyclic changes of the sea level due to glacial freezing and interglacial thawing of polar ice caps. Shelf seas of these periods,—which form the "normal" periods of the Earth's history—were consequently characterized by shallow warm water sedimentation. Moreover, regressions and transgressions of small vertical dimension held sway over wide areas of the shallow shelf sea and the adjoining base-levelled hinterland.

Reef formation during these "normal" periods of the Earth's history was thus governed by an atectonic and non-glacial environment. It must needs have differed from either the present-day reefs, with their glacial control, or the geosynclinal reefs of the past which were tectonically controlled. We believe the Paris Basin to be typical for the sedimentation during these "normal" periods, and a description of some of its reefs is therefore presented.

At present, similar circumstances are to be found only at places where the postglacial eustatic drowning of the shelf has been counteracted, whilst tectonic activity must be low and temperature high. From the literature, it seems probable that in some of the inner zones of the Australian Great Barrier Reef (Fairbridge, 1950), or in some other spots along the Australian coast (Carrigy and Fairbridge, 1954), these qualifications are fulfilled.

FIELD DESCRIPTION

The reefs of the Yonne are found in the Rauracien stage of the Upper Jurassic, a series of purely calcareous strata, between 175 m and 200 m thick, that dip gently northwestward towards the center of the Paris Basin (Abrard, 1950; de Lapparent, 1946). The region is covered by the sheets 96 (Auxerre) and 110 (Clamecy) of the French geological map on 1:80.000. Topographically, the Paris Basin here forms a gently undulating peneplain, ranging in altitude from 200 m to 240 m, whilst the Yonne flows at 140 m at Coulanges and at 130 m at Mailly. On the peneplain the rocks are mostly masked by

Fig. 1. Index map of France, showing location of Yonne reef area. Cross-hatched: Carboniferous and older, lined: Mesozoic + Permian (closer lines where deformed), blank: Cenozoic.

vegetation, the only exposures found being the numerous road cuts and small limestone pits. Along the valleys much better exposures are found, and all classical reef sites (Mailly-le-Château, Mailly-la-Ville, Merry-sur-Yonne and Coulanges-sur-Yonne) are situated on the Yonne river. Here, erosion has cut vertical cliffs up to 40 m high. These cliff faces are not all built up of actual reef limestones, however. In the Mailly-le-Château cliffs, for instance, they form only the lower two-thirds. The upper part of these cliffs is formed by normal sedimentary layers, by oölites and pisolites with a basal bed built up mainly by remains of *Diceras*. Morphologically, there is no difference between the cliff face formed by the reef limestones and that formed by the other sedimentary rocks. Near Merry-sur-Yonne the cliffs, which look like typical reefs (pl. 1), on closer inspection do not contain a reef at all. Their lower two-thirds are formed by fine-grained limestone, with intercalations of thin bands of stromatopores and scarce detrital corals, whilst the upper third is formed by pisolitic limestones, with encrusted detrital coral material. Elsewhere also, reef limestones are found in the small exposures on the peneplain, so there is no reason to suppose that the Yonne had accidentally laid bare an unusual number of the Jurassic reefs.

PLATE 1

Typical cliffs on the Yonne, near Merry-sur-Yonne. They are formed by fine-grained limestone, overlain by pisolites, and do not contain bioherms. Morphologically, there is, however, no difference between the actual reef-cliffs on the Yonne.

All rocks of the Rauracien along the Yonne belong to the *facies coralligène* or *facies récifale* of French authors, that is, to the *reef-rock complex* of MacNeil (1954). Although extremely variable, both laterally and vertically, their color is always whitish, whilst they contain only slight admixtures of terrigeneous matter. They may be porous, compact, or sublithographic, highly fossiliferous or sterile. They may contain bioherms, or only isolated corals in

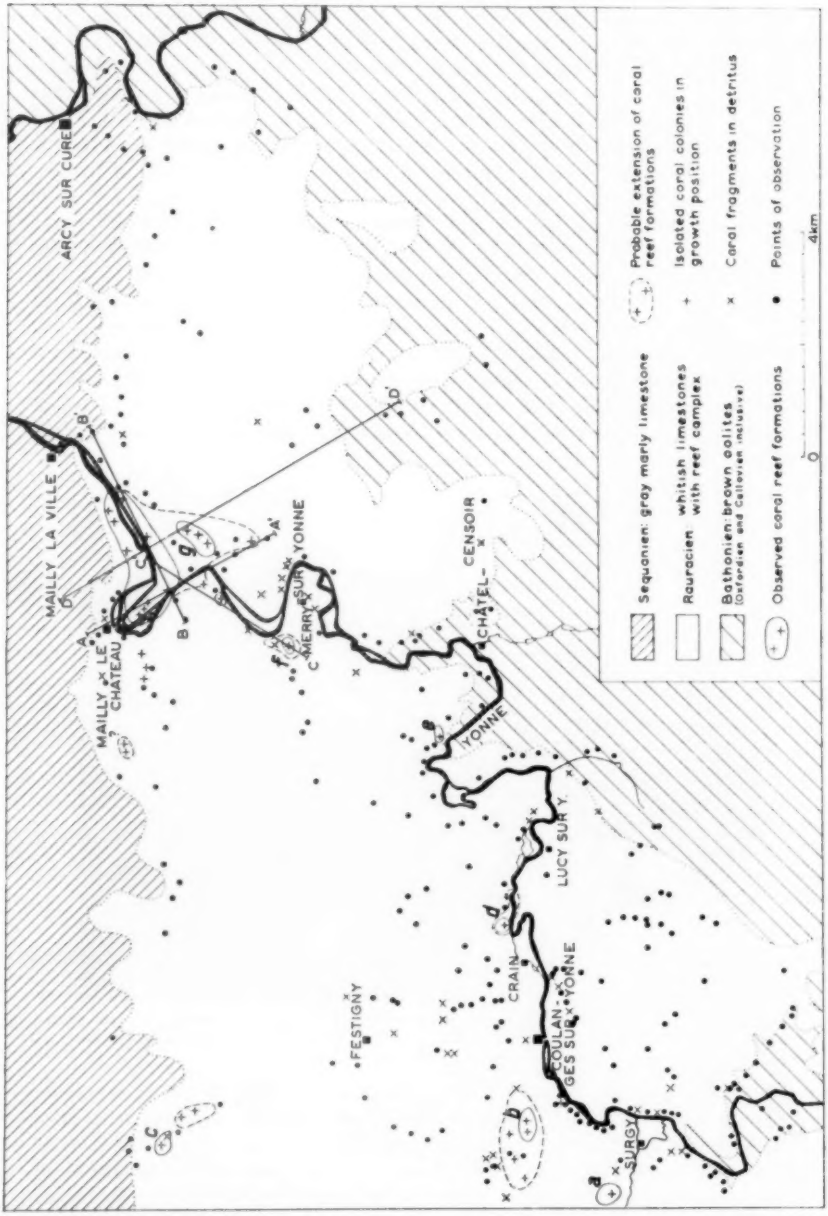

Fig. 2. Map of the reefs in the Rauracien on the Yonne river. Scale 1:120,000.

growth position, but this is exceptional. By far the greater number of these rocks contain no coralligene material, or only a small number of broken specimens. Fine-grained, oölitic and organic-detrital limestones are common, and lateral facies changes are quick everywhere, so there is no regular stratigraphic sequence. In a general way, the lower elements of the series are formed by fine-grained and compact limestones, with stromatopores forming thin biostromes and with scarce detrital coral branches. Corals are mostly found in the middle part of the series, alternating with and also merging into oölitic and fine-grained limestones. Pseudo-oölites and coarse oölites abound in the upper part of the Rauracien. This is no hard and fast rule, however.

Fossils are abundant locally, but the fauna is often poorly developed as to the number of species. Moreover, different species may strongly predominate in different layers. Even the corals in the actual bioherms do not show many different species. This indicates very shallow sedimentation, where small variations in depth, temperature and salinity have temporarily favored the optimal development of a restricted fauna. In a general way, the fauna is found to be connected with certain limestone facies. Thus nerineids are normally found concentrated in thin layers intercalated within a series of fine-grained limestones. It is difficult to say whether these shell beds are *lumachelles* that originated through mechanical concentration, or whether they are real biostromes. *Diceras* commonly occurs in the coarser oölites and in the pisolites, concentrated in a similar way in separate shell beds. Nerineids may occur here too, but they are rolled and encrusted, so they are here in a secondary deposit.

Within this thick zone of limestone of "reef-rock complex" type a surprisingly small amount of corraligene rock is actually found. Bioherms are limited to seven. Of course, several others might be masked by vegetation on the peneplain, but from figure 2 it is clear that bioherms form only a very minor element of the Rauracien. Isolated corals, moreover, are rare too outside the actual bioherms; figure 2 shows all localities where corals have been found, either in growth position (+) or detrital (x).

Bioherms consequently show a haphazard distribution in the Rauracien. They are small and scattered over a wide area. They are not limited to a paleogeographic feature, or to a certain stratigraphic zone of the Rauracien. In the Mailly-le-Château area, they occur almost exclusively in the upper

TABLE 1

Area and Thickness of the Rauracien Bioherms on the Yonne
The bioherms are located by letters a - g on figure 2

Bioherm	Bases	Top	Probable thickness	Observed extension
a	deduced	observed	5 – 10 m	40 x 10 m²
b	deduced	observed	15 – 20 m	300 x 50 m²
c	deduced	deduced	5 – 10 m	150 x 40 m²
d	deduced	deduced	± 15 m	50 x 20 m²
e	observed	deduced	± 10 m	60 x 40 m²
f	deduced	observed	30 – 40 m	100 x 200 m²
g	deduced	observed	> 50 m	2 x 2 km²

part of the Rauracien (fig. 3). Near Coulanges, however, they are found much lower in the Rauracien sequence. Moreover, the thickness of the bioherms is astonishingly small. Where it could be accurately measured, it lies between 5 m and 20 m. Even in the bigger reefs of Merry-sur-Yonne and Mailly-le-Château, extrapolation of the surrounding dips indicates that it does not reach 50 m. Earlier estimates of a thickness of 100 m or more appear to be exaggerated. Consonant with this thinness of the bioherms is the absence of any important syngenetic talus of detrital rocks, exhibiting foreset bedding away from the bioherms. Apparently, the bioherms did not form very marked mounds in the Jurassic sea, and normal sedimentation evidently kept pace with bioherm growth. So, in contrast to earlier descriptions, it must be emphasized that the bioherms really are quite small, both in horizontal and in vertical size. The area, a classical example of reefs of the Jurassic, is not characterized by a large number of bioherms or by much coralligene material, but only by the development of limestones of "reef-rock complex type," with actual reefs very subordinate.

For instance, a single huge reef has been said to reach from Mailly-le-Château for 10 km eastward to the Cure river, but this is certainly not the case. The reef of Mailly-le-Château with a diameter of 2.5 km is by far the largest one in the area. To the east, between the Yonne and the Cure, only normal sedimentary rocks are found. Moreover, the vertical cliff cut by the Cure in the Rauracien south of Arcy-sur-Cure, which superficially resembles a perfect reef, contains a very limited amount of coralligene rock. It is only the fluviatile erosion and subsequent weathering of this cliff that makes it look reef-like, though the rocks are preponderantly normal sediments.

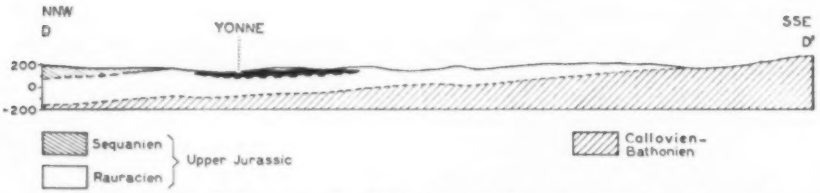

Fig. 3. Section D-D′ across the Rauracien, showing the general position of the Mailly-le-Château reef. Scale 1:60,000. For location see figure 2.

DISCUSSION

Characteristic of the "coralligene" facies of the Rauracien on the Yonne is the extreme development of limestone of the "reef-rock type" and the very small number of actual reefs. A similar situation is encountered in present-day reefs, but here the actual reef is still relatively more important. Even in the "pail-holding-water-structure" of Ladd and Tracey (1949), actual reef-rock is estimated at 5 to 10 percent of the reef mass. In the Jurassic of the Yonne, even this figure is not reached by a long stretch.

There is, moreover, no direct structural relationship between the small reefs on the Yonne and the surrounding series, which is more or less comparable to the present-day "reef-rock complex." This forms a capital distinction

between the recent reefs and the *facies récifale* of the Rauracien. At present, organic debris makes up most of the reef-rock complex, with oölite and fine-grained limestone developing on certain limited areas, protected by coral growth or by coral detritus. In the Jurassic sea, conditions evidently favored the development of oölites and fine-grained limestones over wide areas outside the reefs. Conditions then were probably similar to those existing now in limited areas only, such as the Bahama Bank, shores of the Red Sea, Djakarta (= Batavia) Bay and the like (Rutten, 1949). These rocks are found in present-day "reef-rock complexes" because on recent reefs conditions for their development are favorable too. In the Jurassic seas these conditions were fulfilled over much wider areas and were not restricted to the vicinity of reefs. Therefore, though oölites and fine-grained limestones belong to the "reef-rock complex" of the recent reefs, forming an important part of the reef mass, they are not necessarily restricted to reefs and may develop outside their perimeter, where general conditions of sedimentation were different from those now prevailing.

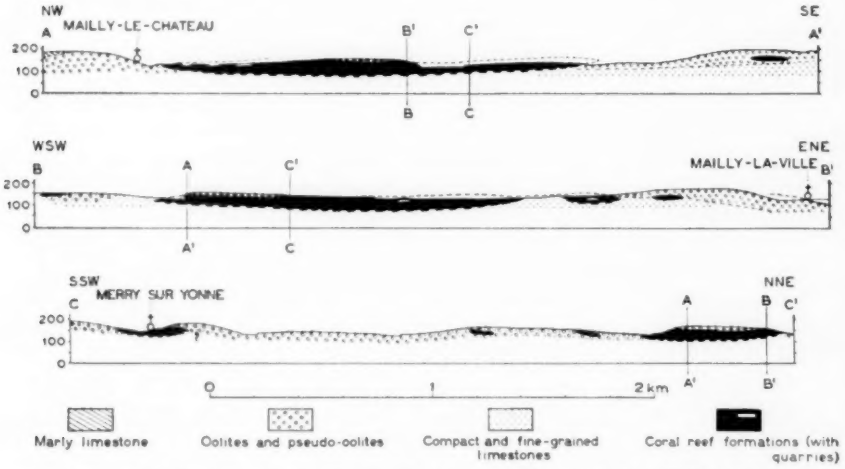

Fig. 4. Sections across the Rauracien reefs on the Yonne. Scale 1:30,000. For location see figure 2.

This seems to be a more general rule. Oölites are common in several stages of the Jurassic in many parts of the Paris Basin and elsewhere, and at best one finds only minor reefs developed locally. Even organic-detrital limestones are a common element of the Jurassic, formed by detritus of a fauna which now is a reef-dwelling fauna solely but which then could form without any direct relation to bioherms or biostromes. In the present seas, with most of the continental shelves drowned by the postglacial eustatic rise of the sea level, apart from scarce banks or bays, lagoons and protected parts of the recent reefs are about the only localities comparable to the wide, warm and shallow epicontinental seas of the Mesozoic.

The contention that "a living reef may supply a limitless amount of reef-rock complex" can consequently not be extrapolated to the atectonic and non-glacial seas of the geological past. On the one hand, several members of the present-day "reef-rock complex" developed independently of reef growth. On the other hand, in typical epicontinental areas like the Paris Basin, the reefs could not attain great thickness or extension, because subsidence during reef growth was slow. The reefs could not outgrow the surrounding normal sedimentation. At the slow pace of epicontinental subsidence, accumulation of the surrounding sediments was able to keep pace with subsidence just as well as a bioherm. The reefs consequently could not form appreciable mounds above the surrounding areas of normal sedimentation.

It follows that the predominance of sedimentary rocks of "reef-rock complex type" in the *facies coralligène* is not due to the presence of actual reefs. This type of sedimentation is more closely linked with the prevailing conditions in the non-glacial and atectonic seas of the Jurassic. The smallness of the reefs which did develop is, moreover, conditioned by the truly epicontinental character of the Paris Basin.

CONCLUSIONS

To conclude, we believe we have demonstrated that the reefs of the Yonne belong to a type different from recent reefs and from reefs in mobile tectonic belts. Characteristic of the "coralligene" facies in epicontinental seas is the independent development of some types of limestone at present belonging to the "reef-rock complex." Development of individual coral growth, of biostromes and bioherms within this so-called coralligene facies may have been very minor and haphazard both in time and in space. The relationship that exists at present between fine-grained limestones, oölites and organic-detrital limestones and reefs, was much less apparent in the geologic past. These rocks then often formed outside reefs and independent of reef development. Present-day and former sedimentation show this difference not because the processes of sedimentation are different but only because a suitable environment for the development of these rocks is very rarely found at present outside reefs, the shelves being drowned by the postglacial eustatic rise of the sea level.

From the tectonically controlled reefs of the past the reefs of the epicontinental seas differ by their slight thickness, a result of the slow and interrupted subsidence characterizing epicontinental areas.

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APPLICATION OF CATION-EXCHANGE REACTIONS TO THE BEIDELLITE OF THE PUTNAM SILT LOAM SOIL*

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ABSTRACT. The titration curve for the reaction of beidellite from the Putnam silt loam soil with KOH can be predicted on the assumption that the beidellite consists of two distinct substrates. The difference in energy of bonding of potassium ion to these two substrates is apparently very small, whereas hydrogen ion is much more firmly bonded to one than to the other. It appears likely that the exchange behavior of a number of clays can be predicted on the basis of a similar assumption that exchange takes place chiefly on two well defined and fairly homogeneous substrates.

INTRODUCTION

The phenomenon of cation exchange on minerals has attracted many workers, and has a voluminous literature. During some attempts to develop a method of describing the process that would permit us to make roughly quantitative predictions concerning exchange reactions between rocks and their contained waters, we discovered that a good "fit" of the KOH titration curve of the beidellite of the Putnam silt loam soil (Marshall and Bergman, 1942) was obtained by assuming that the classical law of mass action holds, and that beidellite behaves like a mixture of two distinct exchange sites.

Although serious doubt has been expressed as to the validity of the primary assumption of the straightforward applicability of the law of mass action (Jenny and Ayers, 1939 and Jarusov, 1937), the closeness of fit of our calculated titration curves to those observed suggested that the method may have specific useful application for workers who are concerned with problems like ours.

There have been many proponents of the simple mass action treatment and they have had success in applying the law of mass action to certain exchange reactions (Vaneslow, 1932). In general they have assumed, for their calculations, that a clay is a reasonably homogeneous substrate, and have used equations of the form

$$XC + Y^+ = YC + X^+$$

which is written to represent the reaction of a Y^+ ion with an X-clay to give Y-clay plus X^+ ion. The mass action relation

$$\frac{(YC)(X^+)}{(XC)(Y^+)} = K_{YC\text{-}XC}$$

provides reasonably constant values of $K_{YC\text{-}XC}$ over a considerable range of X^+ and Y^+ for a number of clays. K values for a number of different cation pairs, although not ideally constant (Davis, 1945), indicate that it is not unlikely that certain clays, at least, can be treated in rough approximation by the same theory that has been so successful in cation fixation on and elution from synthetic resins.

To try our hand at determining an exchange constant for a cation exchange, we first chose Marshall's careful determination of the titration curve for the Putnam beidellite (Marshall and Bergman, 1942, p. 54 and 57). He

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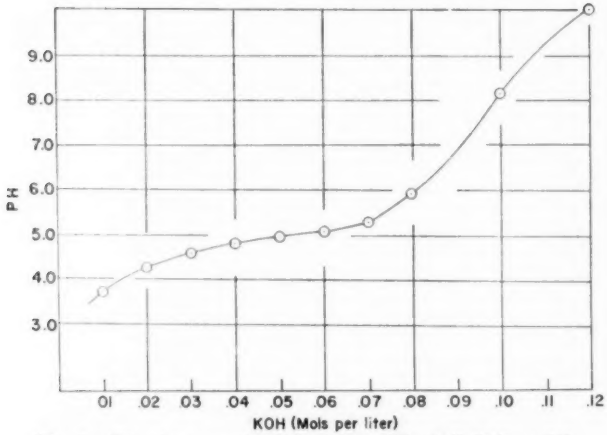

Fig. 1. Titration curve for H-beidellite (Marshall, 1942, p. 54).

converted the clay to the acid, or H-form, then titrated it with potassium hydroxide, measuring pH and K^+ activity electrometrically at various stages of neutralization. A titration curve for an 11.42 weight percent suspension of the clay has been drawn from his data (fig. 1).

EXCHANGE REACTIONS

The titration reaction can be presented:

$$HC + K^+ + OH^- = KC + H_2O$$

where HC represents the acid form of the clay, and KC the potassium-clay complex. From the law of mass action:

$$\frac{a_{KC}}{a_{HC}\, a_{K^+}\, a_{OH^-}} = K_R$$

where a_X is the thermodynamic concentration of each species. Multiplying by the activity product of water:

$$\frac{a_{KC}\, a_{H^+}\, a_{OH^-}}{a_{HC}\, a_{K^+}\, a_{OH^-}} = \frac{a_{KC}\, a_{H^+}}{a_{HC}\, a_{K^+}} = K_R\, K_W = K_{KC-HC} \text{ (exchange constant)}$$

Marshall measured a_{H^+} and a_{K^+}, but the relation between a_{KC} and the concentration of KC, as well as that for a_{HC} and concentration of HC, are not known. However, the ratio $a_{KC/[KC]}$, and that for $a_{HC/[HC]}$, can reasonably be assumed to be about the same (Carleson, 1954, p. 1675). Thus concentrations of the clay complexes probably can be substituted for activities if a ratio of similar species is involved. The exchange equation then becomes:

$$\frac{[KC]\, a_{H^+}}{[HC]\, a_{K^+}} = K_{KC\text{-}HC} \qquad (1)$$

To obtain K, we wrote the stoichiometric relations for the titration:

$$KOH = K^+ + KC \qquad (2)$$

$$HC_0 = HC + KC + C^- \qquad (3)$$

$$K^+ + H^+ = C^- + OH^- \qquad (4)$$

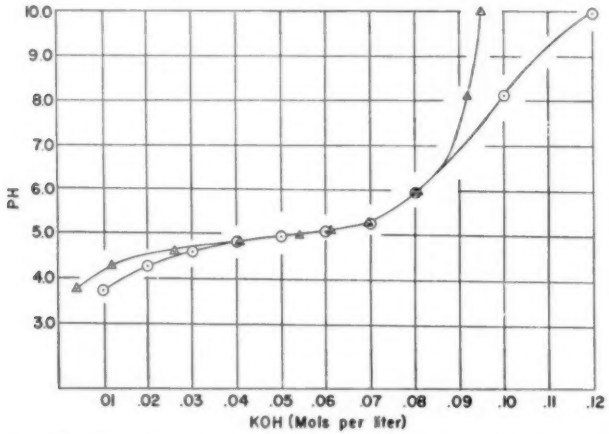

Fig. 2. Calculated (triangles) versus observed (circles) titration curves for Putnam beidellite. Calculation assumes one exchange reaction $\frac{[KC]\, a_{H^+}}{[HC]\, a_{K^+}} = 10^{-2.5}$.

where HC_0 is the original concentration of the clay acid, and C^- the concentration of dissociated clay complex (free sites). Two approximations were used: that the measured activities of the hydrogen and potassium ions are nearly equal to their concentrations at the low ionic strengths involved, and that H^+ and OH^- are insignificant (they are less than $10^{-3.5}$) in equation 4.

Then from Marshall's data for a_{K^+}, a_{H^+} and KOH, values were obtained for the terms of equation 1. The value of HC_0 used in equation 3 was chosen to provide the most nearly constant $K_{KC\text{-}HC}$.

The relation of the calculated titration curve to the observed one, using a "best value" of HC_0 of 0.0835, is shown in figure 2, in which K was taken as $10^{-2.5}$, and the calculation reversed to provide values of KOH from observed a_{H^+} and a_{K^+}. Major deviation from the titration curve occurs both at low and high pH values. Clearly the assumptions are in error, or an exchange equilibrium of the type chosen is not satisfactory, or perhaps not sufficient.

The fact that the observed titration curve "breaks" upward at approximately the same place as the calculated curve, albeit not so sharply, suggested to us that the part of the observed curve beyond pH values of 6 or 7 might represent titration of a second distinct clay acid (HE).

Such an idea is by no means new. Numerous workers, as pointed out by Jarusov (1937) have recognized the presence of two types of bonding of hydrogen ion. He says (p. 286): "The more mobile hydrogen ions are termed 'exchangeable acidity', whereas the less mobile absorbed hydrogen ions are usually characterized as 'hydrolytic soil acidity'." However, mass action expressions based on this view seem not to have been used before.

On the assumption that titration of this second acid does not begin until the first is essentially complete, we determined an equilibrium constant for

it, using the same procedure as before. A value of $10^{-6.7}$ was obtained for $K_{KE\text{-}HE}$. Then it was possible to set up the defining relations for the titration of the two acids simultaneously:

$$\frac{[KC]\, a_{H^+}}{[HC]\, a_{K^+}} = 10^{-2.5}$$

$$\frac{[KE]\, a_{H^+}}{[HE]\, a_{K^+}} = 10^{-6.7}$$

$$KOH = KC + KE + K^+$$
$$HC_0 = HC + KC + C^- = 0.0835$$
$$HE_0 = HE + KE + E^- = 0.03815$$
$$H^+ + K^+ = C^- + E^- + OH^-$$

Then, using again Marshall's values for a_{K^+} and a_{H^+}, the equations were solved for KOH. The calculated curve for the titration is compared with the observed curve in figure 3.

Agreement is good except at the lowest pH values, and deviation there is not great. This residual anomaly probably is related to the effect of H-clay complexes on the glass electrode. The picture has been drawn (Hauser and Reed, 1937) that the original acid clay has a small amount of loosely bound hydrogen ions, and that these, in addition to the free hydrogen ions, contribute to the observed a_{H^+}. They are presumably quickly released during

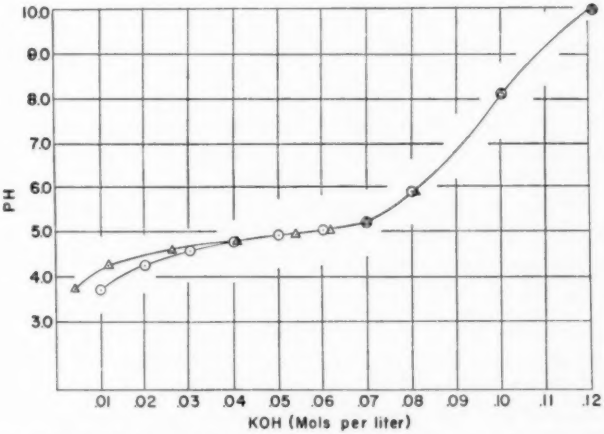

Fig. 3. Calculated (triangles) versus observed (circles) titration curves for Putnam beidellite. Calculation assumes two exchange reactions: $\frac{[KC]\, a_{H^+}}{[HC]\, a_{K^+}} = 10^{-2.5}$ and $\frac{[KE]\, a_{H^+}}{[HE]\, a_{K^+}} = 10^{-6.7}$.

the first part of the titration. If a correction is made for this effect, using an equation of the type:

$$a_{H^+} = [H^+]_{obs.} - [H^+]_{HC}$$

where

$$\log [H^+]_{HC} = k_1 \log [HC] - k_2$$

an essentially perfect fit of the whole titration curve is obtained.

The equations used for the 11.42 weight percent suspension also hold for the curves for 4.567 and 1.142 weight percent suspensions, if HC_0 and HE_0 values proportional to those of the 11.42 weight percent suspension are used.

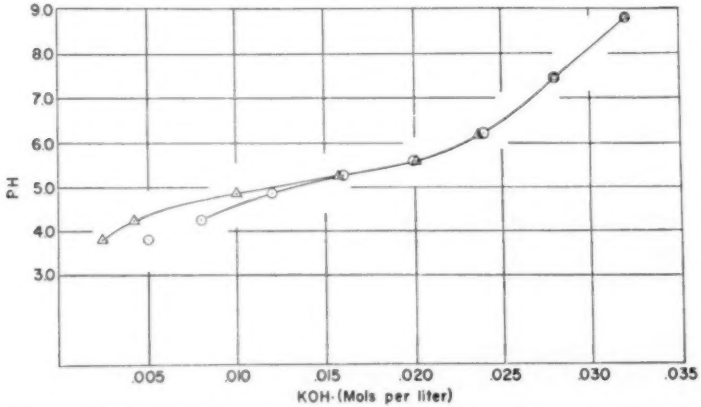

Fig. 4. Calculated (triangles) versus observed (circles) titration curves for "illite." Calculation assumes two exchange reactions:

$$\frac{[KC]\, a_{H^+}}{[HC]\, a_{K^+}} = 10^{-2.4} \text{ and } \frac{[KE]\, a_{H^+}}{[HE]\, a_{K^+}} = 10^{-5.7}.$$

TITRATION OF ILLITE

A similarly satisfactory fit of the titration of a 10 weight percent suspension of illite by KOH (Marshall and Bergman, 1942, p. 56-57) was achieved in the same way. For the illite, with a much lower total exchange capacity, HC_0 is 0.02525, HE_0 is 0.00900, $K_{KC\text{-}HC} = 10^{-2.40}$, and $K_{KE\text{-}HE} = 10^{-5.70}$. The calculated versus the observed curve is shown in figure 4. No correction was made in this case for the effect of the H-colloid on the glass electrode.

DISCUSSION

So far the interpretation of the titration curves has been more or less mechanical, in the sense that the stand has been taken that the beidellite and the illite each behave like a mixture of two independent substrates. It is perhaps particularly noteworthy that the strength of the bonding of the hydrogen ion to these two types of sites (C^- and E^-) is (a) uniform in each site, so that two satisfactory equilibrium constants can be obtained, each related to one kind of site, and (b) the exchange constants for the two sites are markedly

different. For the Putnam beidellite, calculation of the free energy change for the two reactions from the relation $\Delta F^\circ = -RT\ln K$ gives:

$$KC + H^+ \rightarrow HC + K^+ \quad \Delta F^\circ = -3.4 \text{ kcal}$$
$$KE + H^+ \rightarrow HE + K^+ \quad \Delta F^\circ = -9.1 \text{ kcal}$$

Thus the energy loss in exchanging H^+ for K^+ on the E^- substrate is almost three times that for the C^- site.

An even more illuminating relation is that for exchange of H^+ between E^- to C^-, and the corresponding relation for K^+. From the defining equations for the titration it is possible to obtain:

$$\frac{[HE]\ C^-}{[HC]\ E^-} \cong 10^{+4}$$
$$\frac{[KE]\ C^-}{[KC]\ E^-} \cong 10^{0}$$

or

$$HC + E^- \rightarrow HE + C^- \qquad \Delta F^\circ \cong -5.5 \text{ kcal}$$
$$KC + E^- \rightarrow KE + C^- \qquad \Delta F^\circ \cong 0 \text{ kcal}$$

In other words, hydrogen ion is much more strongly bonded to the E^- than to the C^- site, whereas K^+ shows no preference.

The implication is that either two kinds of clay minerals are present, which is unlikely, or perhaps that the C^- sites represent interlayer positions, whereas the E^- sites are "edge" sites. The relative magnitudes of the two exchange capacities, 0.0835 moles for the C^- sites, as opposed to 0.03815 moles for the E^- sites, is in agreement with this concept. Also, it would be expected that H^+ would be more firmly bonded to "edges" than to interlayer positions, whereas K^+, because of its "fit" into the interlayer positions, might be just as firmly bonded as on the "edges."

The presence of two distinct clay acids in a given clay sample suggests interpretations for what has been considered anomalous exchange behavior for the clays. In the first place, it becomes evident that the pH at which cation exchange takes place is an extremely important factor. It is quite conceivable that one clay acid could become saturated with a metal ion while the second acid remained entirely in the H-form. It is also probably possible to prepare a clay with magnesium ions in the exchange positions of the first substrate, and calcium ions in the exchange positions of the second substrate. Under these circumstances displacement of these two ions by a third ion probably would result in a marked change in the ratio of calcium to magnesium during the replacement process. Such an effect may explain the results of Gedroiz (1922) who found that the ratio of calcium to magnesium progressively changed as they were released by treatment of a clay with ammonium chloride.

In addition, it may well be that confusion concerning the relative strength of binding of various cations on the clays may be related to the presence of these two distinct sites. For example, the firmness of binding to "edge" sites may well be related to the charge and radius of the cations, the strength of binding increasing with decreasing values of $\frac{e}{r^2}$, where e is the charge of the ion and r is its radius. On the other hand, strength of binding in inter-

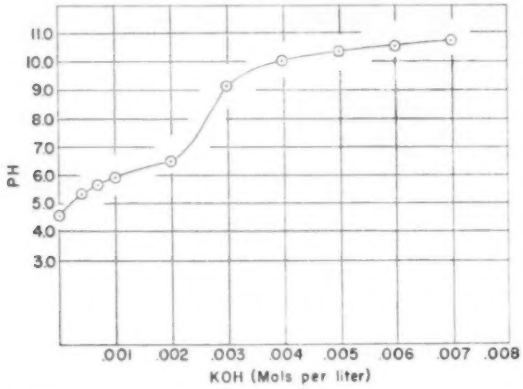

Fig. 5. Observed titration curve for 10.00 weight percent suspension of kaolin (Marshall and Bergman, 1942, p. 56).

layer sites probably is a very different function of radius and charge. Therefore the overall effect observed by any procedure that affects the total exchange capacity of the clay may give a variety of permutations and combinations of these two factors.

The applicability of a single exchange constant for certain cation pairs to many clays and soil colloids may have an explanation in the lack of discrimination of K^+ between C^- and E^- sites. In a $Na^+ - K^+$ exchange, for example, the clay may behave as a single uniform substrate, whereas with cations that discriminate, a single constant will not suffice.

Marshall's titration curve (Marshall and Bergman, 1942, p. 57) for kaolinite (fig. 5) is interesting in light of the interpretation here advanced. The shape of the curve indicates that there are not two distinct clay species —on the other hand no single constant will fit any two consecutive points. This behavior is consistent with an interpretation that hydrogen is held in various sites on kaolinite—on corners, edges, and surfaces—with various degrees of bond strength. Titration with KOH thus should progressively liberate more and more strongly held hydrogens, giving a curve with smaller and smaller exchange constants for each point. Individual "constants" calculated for points show such a decrease with increasing pH.

SUMMARY AND CONCLUSION

It has been shown that titration of the acid form of two clays—a beidellite and an "illite"—can be explained on the assumption that the clays behave as if they were composed of two distinct clay acids. On the basis of this assumption, exchange constants can be determined for the two acids and can be used to duplicate the observed titration curves. It appears likely that the exchange characteristics of some clays can be interpreted on the basis of the presence of two distinct substrates, each with its characteristic binding energy for various cations. If this conclusion is correct it should be possible to treat these clays in the same manner that synthetic-exchange resins have been

studied. If so, it may be possible to make rough predictions of the results when natural waters pass through some clay-bearing rocks.

ACKNOWLEDGMENTS

Sterling Hendricks, of the U. S. Department of Agriculture, and C. E. Marshall, of the University of Missouri, kindly read the manuscript. Both warned of difficulties in the general application of simple mass action exchange relations to clays. Also, we are indebted to our colleague, G. T. Faust, for his helpful discussions of the problem.

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SHORELINE NIPS IN TUFF AT MATSUSHIMA, JAPAN*

K. O. EMERY and HELEN L. FOSTER

ABSTRACT. Well developed intertidal nips in tuff at Matsushima, Japan are believed to have been formed by partial weathering (possibly solution and hydration) followed by removal of the loosened grains by erosional agents. Slow undermining by development of nips has aided in producing the cliffed shores that are characteristic of the region.

DESCRIPTION OF NIPS

Nips occur at the base of seacliffs consisting of tuff, tuffaceous sandstone, and tuffaceous siltstone along the irregular shoreline in the vicinity of Matsushima, Miyagi-ken, Japan, about 350 km north-northeast of Tokyo (fig. 1). They are so characteristic of the landscape in this region that they are even shown on decorative wood carvings (fig. 2). Nips are not common along other coasts of Japan although they are prominent features in limestone cliffs of high islands in the tropics. In the tropics they form intertidal slots that extend from a few centimeters to 5 meters or more into the base of limestone cliffs that border the quiet water of reef flats and lagoons. The nips in limestone are generally attributed to solvent action of sea water owing to biochemical effects of organisms living in the water (Revelle and Emery, in press). Because of the similarity in appearance of the nips in tuff and those in limestone, the authors became interested in learning whether the nips in tuff might also owe their origin to partial solution or weathering by sea water.

The nips at Matsushima are eroded into the Miocene Ajiri formation or its correlatives. In this area the formation is more than 130 meters thick and dips 3° to 10° in most places. Matsushima Bay has an extremely irregular shoreline, and hundreds of small islands and islets are scattered throughout its area (Tomita, 1952). The bay was formed in Pleistocene time by faulting and submergence of the downthrown area. Submergence was followed by slow uplift which is still in progress (Hanzawa and others, 1953, p. 35, 42).

The nips occur along both the seaward- and landward-facing sides of islands and peninsulas, and in some localities are even a little more common on the landward-facing sides. Many small islands, completely encircled by nips, have mushroom shapes (pl. 1-B). The nips extend as much as 1.5 meters into the base of cliffs, and their profiles are somewhat symmetrical within the intertidal zone, especially in places where the tuff is more or less homogeneous. Where the tuff is not homogeneous the nips extend farther into some strata than into others.

Commonly, at the top of the seacliffs, which rise 2 to 10 meters above the water, there is a second overhang (pl. 1-A) the surface of which is harder than the tuff a few centimeters deep; thus, it is a case-hardened surface. This surface originally covered more of the cliff face than at present. The case-hardened material, although somewhat crumbly in places, is harder than the wet tuff of the nips into which a geological pick can be driven to a depth of several centimeters.

The lower half of each nip is partly covered by oysters, whereas the top half has a band of green algae averaging about 45 cm high, above which is

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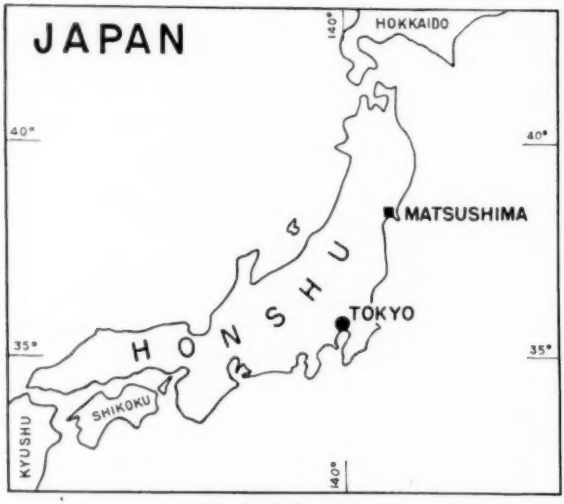

Fig. 1. Index map.

a dark, nearly black band consisting largely of brown algae and averaging about 18 cm high. The top of the black band marks the high tide level (fig. 3). Many small crustaceans live in the cracks and behind thin surface scales of the tuff, especially in the green algal zone. Lichens and moss patchily cover the upper portions of the cliffs. A similar zonation of plants and animals also occurs at the base of cliffs having no nips.

The field observations suggested that the case-hardening was the result of former solution of some part of the tuff by ground water followed by redeposition when the water reached the face of the seacliff and evaporated. Presumably, if fresh water can dissolve part of the tuff, sea water at the base of the cliff might do the same, resulting in weakening of the tuff followed by increased erosion and finally development of a nip.

LABORATORY STUDIES

Petrographic and chemical analyses were made in order to test the field hypothesis of nip development. Thin sections of the tuff show it to be andesitic and composed mostly of pumice with 10 or 15 percent of rock fragments and crystals of plagioclase (andesine), augite, and hornblende. No calcite was detected. Specimens from case-hardened zones contain small quantities of a brown isotropic secondary mineral in cracks and intergranular spaces, and none occurs in specimens from nips; otherwise, no difference was apparent in the thin sections from nips, case-hardened zone, or fresh tuff.

Samples of the rock from the surface of a seacliff were thoroughly dried at 110°C. and leached with concentrated hydrochloric acid at 25°C. for 48 hours. The acid-soluble fraction amounted to 7.5% in the oyster zone, 7.3% in the green algal zone, 8.8% in the black algal zone, and 15.4% in the case-hardened overhang. In order to encompass the extremes in composition,

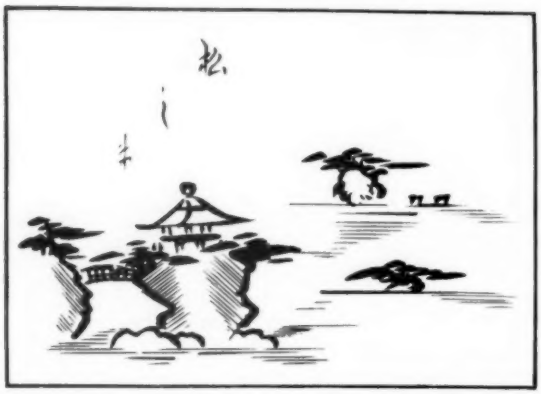

Fig. 2. Carved design on wooden plate from Matsushima showing nips on mainland and around small island.

samples of the green algal zone and of the case-hardened overhang were selected for chemical study. Analyses were made by Japan Inspection Company of Tokyo. The results were recomputed by the author to a basis of 100 percent for total determined constituents (table 1).

The composition of the acid-insoluble fractions (residue of table 1) is similar to that of average andesites as reported by Nockolds (1954), except for lower magnesium and calcium and higher silica and water (above 110°C.). Comparison shows that the acid-soluble fraction differs from the residue in its very small content of silica, but it contains 15 to 40 percent of the original content of all other ions. Much of the material dissolved in acid doubtless represents portions of the rock that had not yet undergone ap-

TABLE 1
Chemical Analyses
(%)

	Green Algal Zone			Case-hardened Zone			49 Average Andesites (Nockolds, 1954)
	Whole sample	Acid soluble	Residue	Whole sample	Acid soluble	Residue	
SiO_2	58.29	0.11	58.18	56.49	0.25	56.24	53.95
$Fe_2O_3+Al_2O_3+TiO_2$	26.23	4.38	21.85	27.00	6.80	20.20	27.75
CaO	4.27	0.65	3.62	4.20	1.72	2.48	7.89
MgO	1.73	0.48	1.25	1.90	0.70	1.20	4.35
Na_2O+K_2O	4.48	1.27	3.21	4.83	1.01	3.82	4.76
P_2O_5	0.19	0.35	4.65	0.28	0.28	5.30	0.28
MnO	0.05			0.08			0.15
H_2O+	4.75			5.22			0.87
Total	100.00	7.24	92.76	100.00	10.76	89.24	100.00

preciable weathering in nature. However, the greater content of the iron, aluminum, and titanium group plus calcium and magnesium in the acid-soluble fraction of the case-hardened zone than in the nip strongly suggests that these ions had been concentrated by nature in the case-hardened zone and removed by nature from the nip zone. Presumably, therefore, they are the ions that have cemented and strengthened the tuff in the case-hardened zone. The small excess percentage ($10.76 - 7.24 = 3.52\%$) of these ions accords with the scarcity of conspicuous cementing material in the thin sections. Similarly, the removal of these ions from the nip zone should have contributed to the observed weakening of the tuff in that zone.

CONCLUSIONS

Field examination indicates that wave erosion alone is not the cause of the nips because the nips are backed by softened and frequently scaly tuff covered in part by attached plants and animals; the intricate shoreline and abundant islands give considerable protection from waves; and finally, the nips are slightly better developed on landward-facing rather than seaward-facing sides of some islands.

These characteristics suggest that, as for nips in limestone, weathering is faster in the intertidal zone than higher on the seacliffs. Weathering of the nips, possibly by solution and hydration, is promoted by the frequent wetting and drying and perhaps is aided by organic acids liberated by attached organisms. Once the grains of tuff have been loosened, they may be dislodged by waves, currents, wind, or activities of animals.

Local case-hardening on the cliff face indicates former strengthening of the surface by deposition of a secondary cementing mineral. Because the cement was probably derived from leaching of the interior of the tuff or of the nearby surface exposures of the tuff, the presence of local case-hardening is compatible with the hypothesis of nip development by intertidal weathering. Similarity of appearance in thin section indicates that only a minor change in the tuff is sufficient to produce either a resistant case-hardened surface or a nip. Chemical analyses show a slight concentration of the iron, aluminum, and titanium group plus calcium and magnesium in the case-hardened zone with respect to the nip zone.

Local weathering and removal of tuff, tuffaceous sandstone, or tuffaceous breccia has elsewhere produced related features. Solution basins in these types of rock occur along the shore or on slightly elevated terraces at Muriwai in New Zealand, on the south end of Guam near Fofos Island, at Hanauma Bay of Oahu, at San Isidro on the Pacific coast of Baja California, Mexico, and probably at many other places.

Undercutting of the lower part of shores by nips, especially where the upper exposed surfaces of tuff have been strengthened by case-hardening, is an important part of the process tending to produce the cliffed shores of Matsushima. The formation of nips by the combined process of weathering followed by removal of the loosened grains by erosional agents is slow but it may eventually result in undercutting so deep that sections of the overhanging cliffs fall and the process begins anew. This may account for the fact that

nips do not occur everywhere along the coast near Matsushima and are developed to differing extents even where conditions seem similar.

PLATE 1

A. Remnants of dark gray case-hardened layer overhanging softer fresh tan tuff. Note presence of algal zones even though nip is not developed.

B. Nips completely encircling a small islet. High tide level is top of black zone.

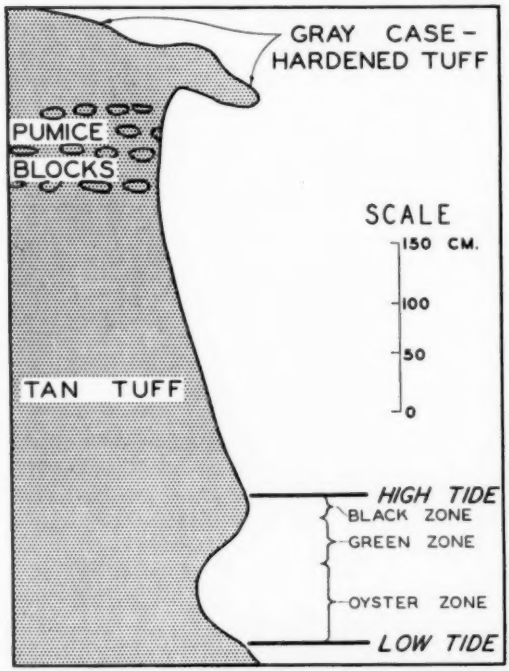

Fig. 3. Sketch of typical nip, showing zonation and relationship to tide levels.

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REVIEWS

Quantitative Spectrochemical Analysis of Silicates; by L. H. AHRENS. P. 122; 27 figs., 7 pls. Cambridge, Mass., 1955 (Addison-Wesley Publishing Company, $3.75).—The book—consisting of nine chapters, a bibliography, and a subject index—describes the use of the d. c. arc for the spectrographic analysis of silicate minerals, rocks, soils, and meteorites.

The method is described with sufficient detail for the inexperienced as well as the experienced spectrographer to complete an analysis. The tables list analytical data on electrode purity, granite, diabase, alkalies in metamorphic rocks, and many others which will be of interest to spectrographers, geologists, geochemists, and other scientists.

The main analytical scheme of three to seven steps for the determination of 35 elements is a definite contribution to the field of nonmetal, emission spectroscopy and will probably find its way into many spectrographic libraries. The book should be inspected by spectrographers in the metals field with the possibility of converting the metal samples to oxides where special cases are involved (segregation, etc.) in making an analysis.

The author does not discuss stroboscopic and intermittency effects as possible contributors to errors in spectrographic analyses when the sector wheel is used to vary the line intensities on the spectrographic plate. The author recommends the use of the sector wheel and, therefore, should have issued a warning, especially for inexperienced spectrographers.

The author's treatment of the spectrochemical errors will be of particular interest to those spectrographers supplying analytical data as well as to the scientists who apply the data to specific problems. If the dispersion of the error is small (less than about 15 percent), a normal distribution may be assumed; otherwise the spectrochemical error distribution is of the lognormal type where the geometric mean of a series of replicate determinations replaces the arithmetic mean for the calculation of the line intensities or intensity ratios. Statisticians will be interested in the discussion and examples of spectrographic error.

This volume is the third of a series of well received books by the same author dealing with spectroscopy. It is recommended to spectrographers, geologists, metallurgists, and all others who have inorganic analytical problems.

CLAUDE L. WARING

The Principles of Chemical Weathering; by W. D. KELLER. P. 88; 21 figs. Columbia, Missouri, 1955 (Lucas Brothers Publishers, $2.75).—Is chemical weathering the geologic process most important to man? Professor Keller, in reviewing the geologic (sensu vasto) aspects of chemical weathering, sets out to show that there can be no doubt of it. Professor Keller is preëminently qualified to discuss weathering and he presents a strong case for his viewpoint. His abundant and almost disarming knowledge and enthusiasm will provide a genuine stimulation for the undergraduate student in Earth science.

Keller suggests one novel concept to explain the mechanism of pseudomorphism that deserves widespread consideration. This mechanism "is determined by the energy of the attacking reactant relative to the bonding

energy (stability) of the host." Thus he explains the phenomenon of nonstructural pseudomorphism in contrast to the type in which the guest assumes the internal structure of the host. By an elaboration of this mechanism Keller reconciles the widely divergent views on mineral transformation, whether in the liquid state or in the solid state.

Unfortunately, the book, which is reproduced by photo-offset lithography, is marred by misspellings and misleading definitions. Its usefulness will not be impaired, however, if the user proceeds with a cautious and critical approach.

KURT SERVOS

Structure Contour Map of the Montana Plains; by C. E. DOBBIN and C. E. ERDMAN. U. S. Geol. Survey, Oil Investigations Map OM 178, 3d revision. Washington, D. C., 1955.—This useful map, first published in 1932, was reprinted with revisions in 1935 and 1946. Now it is again brought up to date, to present in graphic form a wealth of geologic information, much of it won by field studies since the Second World War. The new edition is available in two forms: Map OM 178-A, scale 1:500,000, on a sheet 48 by 72 inches; and Map OM 178-B, scale 1:1,000,000, measuring 24 by 36 inches. The larger map is intended for the wall, the smaller map for desk or field use.

Structure contours, in red, with 250-feet interval and drawn on the base of the Colorado shale (actual or restored) picture an impressive array of domes, anticlines, and structural basins. But the map presents much more than its title implies; it is in fact a tectonic map, showing traces of major faults such as the Lewis and Beartooth thrusts, patterns of en echelon belts of faulting such as the Lake Basin, and lithologic units such as the Boulder batholith, the Crazy Mountains stocks, and many others. Contours are continuous over the plains and the outlying foothills units, and end irregularly along the steep mountain front. Tectonic and lithologic symbols are carried somewhat farther west, but the high mountain belt is omitted. Space in the western margin of the sheet is used for explanation of symbols and for a generalized stratigraphic diagram.

Careful study of the map, with sample checking for accuracy, brings conviction that someone spent countless hours in digesting source materials and in laborious plotting, to give a faithful structural picture of Montana east of the Rockies. We are much in debt to the authors of this tectonic map.

CHESTER R. LONGWELL

The Geology of the Cuyuna District, Minnesota; by F. F. GROUT and J. F. WOLFF, SR. Minnesota Geological Survey Bull. 36. P. xiii, 144; 24 figs., 6 pls. Minneapolis, 1955 (University of Minnesota Press, $3.00).—The results of geologic investigations extending over a period of 37 years are presented in this first major report on the Cuyuna District since 1919. The book is divided into seven chapters. The first three, written jointly by both authors, include an introduction, a general setting, and a description of the Cuyuna ores; the fourth includes both Grout's account of the stratigraphic succession of the main or North Range as indicated by recent drilling and Wolff's description of the geologic map and cross sections; the fifth and sixth, again

by both authors, cover a discussion of outlying areas and a correlation of the Cuyuna stratigraphic units with those of the Mesabi Range; and finally, the seventh chapter by Grout is a brief discussion of the low-grade manganese ores of the Cuyuna District.

The lack of full agreement on the stratigraphic succession within the Cuyuna district has been due to the thick glacial deposits, the total lack of bedrock exposures, and the complex structure. Basing their work on a great deal of subsurface data the authors recognize in the Cuyuna district a series of three stratigraphic units: a lower unit of quartzite and quartz slate, a middle unit of cherty and slaty iron formation, and an upper unit of slate which also contains a thin iron-bearing member. The middle iron-bearing unit forms the North Cuyuna Range, and the upper unit the South Cuyuna Range. These relationships are shown by the colored geologic map and cross sections, prepared by Wolff on the basis of drill records, geophysical surveys, and exposures in mines, many of which are now inaccessible. Some drilling financed by the University of Minnesota has also contributed new stratigraphic information supporting the suggestion that the iron-bearing unit of the North Range can be divided into four members which are analogous to the four members of the Mesabi Biwabik formation. This relationship, together with the overall similarity of the threefold stratigraphic sequence of the Cuyuna district with that of the Mesabi Range, has led the authors to correlate the three Cuyuna formations with the Pokegama, Biwabik, and Virginia formations of the Mesabi Range. The authors, therefore, use these terms in the Cuyuna district but point out some of the difficulties of this correlation and suggest many problems yet to be solved.

The section on the origin of the Cuyuna iron ores is disappointingly short. It is not, however, an important part of the report, being overshadowed by the more important sections on the development of the stratigraphic succession, the correlations with the Mesabi Range, and the compilation of the geologic map. The report should be a helpful guide for geologists working under the handicap of a thick drift cover in an area of complex geology.

FRANK G. LESURE

Notes en rapport avec l'hydrogéologie des roches calcareuses; by IVAN DE RADITZKY D'OSTROWICK. P. 199; 59 figs. Dinant, Belgium, 1953 (L. Bourdeaux-Capelle, paper bound, Belgian francs 160).—This book is a reprint of a monograph on the hydrogeology, or ground-water geology, of calcareous rocks. The monograph first appeared in the Bulletin de la Société Royale d'Etudes Géologiques et Archéologiques, v. 15, 1953.

Introductory sections on the origin of rocks, calcareous rocks of Belgium, permeability, and the hydrologic cycle prepare the nonspecialist for the more technical parts of the book. The origin of limestone caverns and related topics such as the solution of limestone, deep aquifers in limestone, forms produced by the solution of limestone, and cave deposits are discussed in the first half of the book. The theories of W. M. Davis and J H. Bretz concerning the origin of caverns are reviewed at length. Special attention is devoted to the criteria used to establish the phreatic origin of various features found in limestone

caverns, and the conclusion is reached that most of the criteria are not applicable to caverns in Belgium. The author does, however, admit that they are probably valid in the United States where they have been used and concludes that the matter needs further study in Belgium.

The remainder of the book is devoted to the relation of such diverse topics as mineral deposits, ground-water resources, dam foundations, and water pollution to the circulation of ground water in calcareous rocks. Despite a large number of caves and other solution features described from Belgium, the book is not provincial in scope. Many of the local examples are compared with features found in England, France, and the United States; moreover the important literature from these countries is covered well in the text. This book should be useful to geologists not only as a general reference on the ground-water geology of calcareous rocks but also as a collection of valuable field observations by a scientist whose interest in the subject spans four decades. STANLEY N. DAVIS

Prehistory and Pleistocene Geology in Cyrenaican Libya; by C. B. M. McBURNEY and R. W. HEY. P. xii, 315; 40 figs., 16 pls. New York, 1955 (Cambridge University Press, $10.00).—This interesting report illustrates the advantage of conducting geologic and archeologic field studies simultaneously in the same district, over the more common practice of "going it alone" —and missing critical parts of the evidence. For although the book consists mainly of two separate and very competent contributions, by the geologist (Hey) and the prehistorian (McBurney), each clearly reflects the benefits conferred by collaboration. The result is a correlation of late Pleistocene events and cultures that could hardly have been achieved if the two studies had been made at widely different times.

The field work, organized from the University of Cambridge and carried out in 1947, 1948, and 1952, covered the coast of Cyrenaica between Benghazi and Derna. The Pleistocene deposits examined include alluvium, colluvium, eolianites, and tufas, but by far the most interesting features dealt with are emerged marine strandlines accompanied by thin beach gravels. Some of these features were found to have good continuity through distances up to 110 km. The sequence consists of six units, ranging from more than 140 m down to 6 m above present sea level, and increasing in age with increasing altitude. Marine deposits near Benghazi demonstrate, however, that a sea level lower than today's intervened between numbers 2 and 3 measured from the lowest. Hey is of the opinion that the whole sequence of strandlines is almost solely the result of changes of sea level; only local and minor crustal deformation seems to have occurred.

He reaches the conclusion (which will startle some students of the Pleistocene) that the Cyrenaican strandlines are the correlatives of the "classical" sequence of Depéret elsewhere in the Mediterranean region, and that the lowest (that at 6 m) probably dates from the last interglacial age. Inferences as to climatic conditions near the end of the sequence are drawn from colluvium, eolianites, and other terrestrial sediments whose stratigraphic relations to strandlines were established.

On the archeologic side a cultural sequence was reconstructed from evidence locally rich: lower, middle, and upper Paleolithic cultures were identified, as well as two Neolithic cultures, contemporaneous or nearly so. Specifically the sequence includes: (1) A hand-axe industry dating from a time of greater rainfall, and a late Acheulean culture. In the middle Paleolithic, a flake industry dating from the lowest strandline (last interglacial), a Levalloiso-Mousterian correlated with a moist period (Würm I?), and a Levallois-like culture identified with a cold period (Würm II?) of frost-wedging and accelerated mass-wasting. (2) In the upper Paleolithic, two successive blade industries. The lower one suggests the lower Aurignacian of the Levant and the Capsian of Tunisia, and is related to a cold dry climate and to a radiocarbon date of ca. 16,000 yr B.P. The upper one is dated at 10,000-7,000 yr B.P.

The contributions by both authors are fully documented and are reasoned deliberately and with caution. One could wish that a similar double-barreled weapon could be leveled at other critical segments of the Mediterranean coast.

RICHARD FOSTER FLINT

The Life in the Sea; by RALPH BUCHSBAUM. Condon Lectures, Oregon State System of Higher Education. P. x, 101; 9 figs., 35 pls. Eugene, 1955 (University of Oregon Press, $1.50).—Although another popular book about the sea may appear superfluous to the reading public, the value of Dr. Buchsbaum's book lies not so much in his information about the sea as in his historical summary of man's efforts to understand the sea and its indigenous populations. The historical section (the first chapter) includes fully one-third of the book; the second chapter deals with the animal communities in the sea, with emphasis on the littoral for the benefit of the casual observer; and the third chapter summarizes some of the current research and emphasizes that which is economically important.

The historical summary begins with a mention of prehistoric cultures living near the sea and terminates with the 20th century spurt in the building of marine stations. Within this framework the contributions of the Greek philosophers and the effects of the *Challenger* expedition (1872-1876) are particularly accentuated. The summary as a whole is a vehicle for a discussion of two points: western man's gradually diminishing fear of the sea, and the stagnation which may be produced by an authoritarian point of view which prevents the free exchange of ideas.

Two points come to mind concerning the history of deep ocean sounding. If one must pinpoint the first efforts at deep water sounding, Columbus' efforts should be recognized since they occurred nearly 30 years before Magellan's. Columbus attempted a sounding in the Sargasso Sea on September 20, 1492, and found no bottom at 200 fathoms—the bottom at that point is circa 2292 fathoms (Morison, 1942). Also at the time of these lectures the deepest modern recorded sounding had been made by H.M.S. *Challenger* in the Mariana Trench on June 14, 1951, a sounding of 5940 fathoms (Carruthers and Lawford, 1952).

In chapter 2 the concept of the community and of the food chain in the ocean is outlined, and a summary of the physical and biological factors which affect the marine populations is included. This section is slightly disorganized and contains more than a few errors in terminology. The concept of communities is illustrated by a description of the zonation of rocky shores as well as a description of the differences between rocky, sandy, and muddy shore populations. A few drawings are included that are based on works of other authors, some of which are not listed in the references.

Chapter 3 contains short summaries of some current research in oceanography, such as: bottom sediment analyses, methods of estimating temperature of ancient seas, plankton analyses leading to the discovery of indicators of water masses, and productivity studies. The author dwells briefly on those problems which directly influence our economy.

The purpose of the Condon lectures is "to interpret the results of significant scientific research to the non-specialist." This Dr. Buchsbaum has done with the addition of over 60 photographs, more than one-half of which are his own. This book will be especially helpful to the seaside vacationer who is interested in marine biology—its growth and its future.

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